AD-A019 703

MOUNTAIN AND GLACIER TERRAIN STUDY AND RELATED INVESTIGATIONS IN THE JUNEAU ICEFIELD REGION, ALASKA-CANADA

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Foundation for Glacier and Environmental Research

Prepared for:

Army Research Office

September 1975

DISTRIBUTED BY:

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U. S. DEPARTMENT OF COMMERCE

MOUNTAIN AND GLACIER TERRAIN STUDY AND RELATED INVESTIGATIONS IN THE JUNEAU ICEFIELD REGION, Alaska-Canada

by
Maynard M. Miller

and

Research Affiliates of the
Juneau Icefield Research Program

FINAL REPORT

U.S. Army Research Office-Durham

Grants No: DA-ARO-D-31-124-71-G120
DA-ARO-D-31-124-72-G193
DA-ARO-D-31-124-[illegible]-G185

Foundation for Glacier and Environmental Research
Pacific Science Center
Seattle, Washington 98109

Monograph Series

May 1975

Unclassified

Security Classification

DOCUMENT CONTROL DATA - R & D

(Security classification of title, body of abstract and indexing annotation must be entered when the overall report is classified)

1. ORIGINATING ACTIVITY (Corporate author) Foundation for Glacier and Environmental Research 4533-19th Ave. N.E. Seattle, Washington 98105	2a. REPORT SECURITY CLASSIFICATION Unclassified 2b. GROUP NA
3. REPORT TITLE MOUNTAIN AND GLACIER TERRAIN STUDY AND RELATED INVESTIGATIONS IN THE JUNEAU ICEFIELD REGION, ALASKA-CANADA	
4. DESCRIPTIVE NOTES (Type of report and inclusive dates) Final (covering period 1971-1975)	
5. AUTHOR(S) (First name, middle initial, last name) Maynard M. Miller	
6. REPORT DATE May 1975 (rev. September 1975)	7a. TOTAL NO. OF PAGES 263 / 7b. NO. OF REFS
8a. CONTRACT OR GRANT NO. DA-ARO-D-31-124-71-G120 b. PROJECT NO. DA-ARO-D-31-124-72-G193 DA-ARO-D-31-lar-73-G185 c. d.	9a. ORIGINATOR'S REPORT NUMBER(S) Unnumbered monograph series, Foundation for Glacier Research 9b. OTHER REPORT NO(S) (Any other numbers that may be assigned this report)
10. DISTRIBUTION STATEMENT Approved for public release; distribution unlimited.	
11. SUPPLEMENTARY NOTES	12. SPONSORING MILITARY ACTIVITY U. S. Army Research Office-Durham Box CM, Duke Station Durham, North Carolina 27706

13. ABSTRACT

The investigations cover the years 1971-74 on the Juneau Icefield, Alaska (the Taku District) and in the Atlin, B.C. region. Significant advantage has been gained by the availability of field data from the Juneau Icefield Research Program since 1946. With our aim the clarification of the process factor in arctic and mountain terrains, emphasis is on glacial, glacio-fluvial and periglacial landforms. Because the effectiveness of process stress is controlled by geologic structure and time, and because this stress represents a force which combines mass and acceleration, the total mass/energy continuum has to be considered. In this, attention is paid to energy distribution in the system and to its changes...i.e., the entropy of the terrain being studied. The analyses are made from data in the disciplines of climatology, glaciology, hydrology, geophysics, continuum mechanics and periglacial and glacial geology.

The Juneau Icefield serves as a continental divide, west of which are 60 miles (100 km) of drainage to the Pacific Ocean. This is compared to 2300 miles (3800 km) of inland drainage northward down the Yukon River to the Bering Sea. The proximity of the Gulf of Alaska results in high-intensity coastal storms which bring much moisture and energy into the system. This research, therefore, has glacio-climatological emphasis.

Data from 14 meteorological stations are analyzed along a 200-mile transect between Juneau, Alaska and Whitehorse, Yukon. A gradient of continentality is revealed, with a "sun-line" identified in a transitional zone 10 miles (16 kms) northeast of the ice divide. Between here and the coast, high humidity prevails, with excessive summer melting on the Alaskan névés, producing rilled and sun-cupped surfaces alternately destroyed and re-formed by rain and sun. Inland, evaporation adds to ablation and pro-

(cont.)

duces highly roughened surfaces. Below the nêvé-line, melt-water flows by supraglacial streams into moulins and moves by intra-granular permeation into englacial channels.

Allied with the continentality gradient is a rise in nêvé-lines, which are 2000 feet (650 m) higher on the continental flank. Since 1945, these have varied as much as 1000 feet (300 m). Regional cirque and berm levels and tank and tor topography are investigated regarding former glaciation limits and Pleistocene nêvé-lines. A 9-fold cirque system in the Taku District compares to 5 levels in the Atlin area, corroborating that glacial processes on the coastal flank far exceed those inland. The continental sector, however, has experienced stronger development of periglacial features (palsas, patterned ground and rock glaciers) during the Wisconsinan. Permafrost is still common above 5500 feet (1700 m).

Our glaciological research emphasizes mass balance and the deformation of ice masses such as the Lemon and Cathedral Glaciers which, in this study, are coastal and inland prototypes. Firn stratigraphy shows a lowering of maximum accumulation levels in the past 15 years...from up to 7000 feet (2100 m) in the 1940's and 50's down to 3000 feet (900m) in the 1960's and 70's. The data verify that with warming phases there is a coastward shift in the Arctic Front coincident with upward migration of transient snow-lines. Measured mass balances corroborate climatic trends in the regional meteorological records at Juneau and Whitehorse over the years 1907-1974.

Glacier regimes are also evaluated with respect to Little Ice Age and Neoglacial histories. Lichenometric and palynological methods have been fruitful, assisted by isotope dating of organics in bogs and till-buried intra-glacial forests. The regional glacial landform sequence provides a Quaternary chronology. In this, the observed minor cycles, involving 20- to 30-year and 80 to 90-year shifts in storm paths, are found to manifest the nature of much larger displacements in the mean position of cyclonic storm centers during the Holocene. During periods of cooling, the circumpolar vortex retracts and storm paths move inland. With this shift in peripheral storms, there is a change to southwesterly winds bringing added moisture inland from the Pacific.

A relationship between glacier fluctuations and long-term mean positions of the dominant North American tropospheric ridge seems to be coincident with a 45-year cooling trend (90-year cycle) expected to bottom out about 2000 A.D. The solar control mechanism is examined with consideration given to zonal index changes in the circumpolar vortex. Intensity of the continentality gradient appears to vary as changes in pressure take place in the tropospheric westerlies, apparently in response to variations in electromagnetic and corpuscular radiation. Additional proof of this indicated correlation will be sought in our continuing research. The important controls of elevation and geographic position have also been recognized in interpreting changes in glacier regimes and their hydrological consequence.

The evolution of lesser terrain features, such as self-spilling lakes, glacier caves, fluted channels, wave bulges, ogives, firn folds, bedrock thresholds and phenomena associated with avalanches and Jokulhlaups are also discussed. Much is preliminary to further investigations in this region which, by its unique geographical location, is so sensitive to the process factor. But even in the incomplete aspects of this study, the self-regulation of natural systems constantly striving for a more even distribution of energy is well illustrated. Overall this terrain must remain in high relief because of the combined endogenetic and exogenetic processes involved. The interplay includes not only subaerial stresses connected with climate, but crustal deformations in an earthquake belt which remains in disequilibrium and at a low entropy level. The resulting imbalances in landscape are dynamic, because structural controls and incessantly strong energy inputs, both periodic and aperiodic, produce all the characteristics of a most probable state.

These investigations illustrate fundamental principles of terrain evolution. The principles involved not only are imporant in this part of the sub-arctic but also in regions of the high arctic of North America where so much developmental activity is now centered.

14 KEY WORDS	LINK A		LINK B		LINK C	
	ROLE	WT	ROLE	WT	ROLE	WT
Arctic and mountain terrains						
Glacier research						
Glaciology						
Juneau Icefield Research Program						
Juneau Icefield						
Arctic geomorphology						
Glacio-geomorphology						
Glacio-hydrology						
Glacio-meteorology						
Arctic-climatology						
Alaskan coastal geomorphology						
Glacier continuum mechanics						
Structural glaciology						
Taku District						
Atlin District						
Glacier regimens						
Snow and ice terrains						
Pleistocene research						
Alaskan glaciers						
A						

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* THE FINDINGS IN THIS REPORT ARE NOT TO BE CONSTRUED AS AN OFFICIAL DEPARTMENT OF THE ARMY POSITION, UNLESS SO DESIGNATED BY OTHER AUTHORIZED DOCUMENTS

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PART I INTRODUCTION AND OVERVIEW

The investigations reported here are part of the Juneau Icefield Research Program (JIRP) which since its inception in 1946 has aimed at a long-range total environment study of the fifth largest icefield in North America (Fig. 1). The glaciers in this region lie between 58° and 60° N Latitude and so technically may be referred to as Sub-Arctic. But as every 1000 feet (330 m) of elevation rise in the Alaska-Boundary Range is climatologically equivalent to pushing the climate 300 miles north, at the névé crest of the Juneau Icefield (ca. 8000 ft, 2424 m) climatic conditions are high Arctic in character...i.e., even in summer being essentially Polar. Because of this elevation range, the larger Juneau Icefield glaciers are thermo-physically polythermal (Miller, 1975c). Thus they are a most sensitive indicator of long-term climatic change (Lawrence, 1950; Miller, 1964a) Furthermore, the Juneau Icefield, situated as it is in the interaction zone between cyclonic and anti-cyclonic pressure cells at the northern edge of the Gulf of Alaska, is uniquely sensitive both to short-term and secular climatic trends and, hence, serves as a global prototype for long-period studies of glacier and snow-covered regimes (Smithsonian Science Project abstracts). Some of the long-term aspects of the JIRP study are now in their 30th consecutive year (Field and Miller, 1951).

This report concerns investigations conducted during the summers of 1971, 1972 and 1973, as supported by grants to the Foundation for Glacier and Environmental Research from the U.S. Army Research Office-Durham (Grant Numbers: DA-ARO-D-31-124-71-G120; DA-ARO-D-31-124-72-G193; DA-ARO-D-31-124-73-G185).

Purpose and Scope of the Research

The basic aim of the Army Research Office support has been to assist those aspects of the JIRP activity which involve investigation of mountain and arctic terrains on the Juneau Icefield and its peripheral maritime and continental regions, with emphasis on snow and ice regimes and the effects on these regimes of orographical and climatological factors. In this report, special attention is given to seasonal and longer-term regimes relating to factors of climatic change and to the geographical and geological aspects of elevation and of the associated maritime and continental environments.

Thus, a key purpose of the field work has been to obtain a systematic series of environmentally related measurements in the glaciological sectors of the Juneau Icefield south and west of the Alaskan border and in the adjoining non-glacierized sectors of the Atlin Lake region in northern British Columbia and the Yukon Territory. These measurements concern the extent and configuration of the icefield surface and its glaciological characteristics as they affect the nature of this terrain. The practical relationship of glacial and periglacial topography and subsurface structure to the trafficability of tracked vehicles and to ski and foot travel in remote mountain and Arctic regions is important to know. But through investigation of this climatically sensitive region, a deeper purpose is revealed... i.e., to determine more precisely the relationship of terrain changes to broader climatic trends. This includes incorporating relevant information from other Arctic regions concerning the intensity, depth and character of permafrost and the summer "active zones", and the volume and aerial variation of drift ice and pack-ice in Arctic waters, the latter being equally sensitive indicators of short-term climatic trends.

The scope of this study thus involves the intertwining of meteorology, glaciology and surveying and photogrammetry at a network of field stations which are aligned on a transect across the Juneau Icefield from the maritime coast of the Boundary Range to the dry interior of its flanking continental region (Fig. 2). The investigational scope also embraces hydrological, geophysical and continuum mechanics studies at selected sites where measurements of englacial change can be made in relation to both seasonal and annual climatic perterbations and to the morphogenesis of snow and ice features. Also to be considered in this report is the work done on glacier terminal behavior patterns during Neoglacial time, on Pleistocene terrain features and landforms and finally in the chronology of the Quaternary pertaining to this region with emphasis on the Holocene.

Emphasis on Process Intensity and Terrain Evaluation

An underlying concern in these studies is on the process factor in the landform equation stated with terrain as a function of the geologic material involved (e.g., snow, ice or permafrost) plus the effects of process type and intensity through time. In this the following process oriented research has been conducted: (1) synoptic regional meteorology; (2) local glacio-climatology; (3) glacio-hydrology including the study of ice-impounded water bodies and their sudden break outs or hydrological surges (Jokuhlhaups); (4) study of the causes of moat formation on glacier margins; (5) gaging of out-flow in glacial streams; (6) water-stage recording of the changing surface levels of pro-glacial lakes; (7) exploration and mapping of glacier caves as hydrological reservoirs; (8) investigation of ice avalanches; (9) recording of glacier micro-seisms using portable seismographs; (10) key surveys of surface ice movement on across-glacier and longitudinal profiles; (11) structural glaciological studies and their relationship to surface forms; (12) surface snow-cup and roughness studies; (13) research on englacial temperature characteristics and allied diagenetic ice development; (14) review of patterns of glacier terminal fluctuations in the area; (15) research on the processes of palsa development in periglacial situations; (16) a long-term survey of rock glacier flow and allied surface terraces; and (17) study of the morphogenetic chronology of key geomorphic terrain features of erosional and depositional form and their structural and lithologic controls. This includes such major terrain features as tandem cirques, berm sequences, esker complexes and crevasse-fillings and moraine patterns.

Another phase of the program is the investigation by geophysical and surveying techniques of the mechanics and causes of periodic glacier variations, a study closely allied to earlier research efforts of JIRP. This includes their relation to secular climatic trends and the differentiation of anomalous ice pulsations apparently related to dynamic instabilities which develop in glacier systems when their self-regulating capabilities exceed the boundary limits imposed by load and temperature parameters in the system. Part of this concerns the relationship to changing snow-pack thicknesses and snow-cover extent and the accumulation characteristics of mountain névés.* In a broader context, it relates as well to thickness and area changes of arctic sea ice in consequence of global climatic trends. The specific investigations conducted in 1971 through 1973 include interpretive study of short-term changes on several representative glaciers of the Juneau Icefield region and the study of longer-term snowfall and ablation changes affecting the upper nourishment surfaces of these prototype glaciers.

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* Névé with an areal connotation, as the permanent snow-field (firn-field) accumulation area as limited by the névé-line late in the summer.

Sixteen specific glaciers were the loci of process studies. These include the Ptarmigan, Lemon, Mendenhall, Herbert, Taku, Hole-in-Wall, East and West Twin, the Vaughan Lewis, Gilkey, Bucher, Talsekwe, Llewellyn, Williston, Hoboe and Cathedral Glaciers. Permanent research station facilities exist on the Lemon-Ptarmigan Glacier system and on the Taku, Llewellyn, upper Bucher and Cathedral Glaciers, with the result that greatest attention was concentrated in these areas. Of these ice masses, the Taku and Hole-in-Wall Glaciers are advancing (as they have since 1894), the Lemon-Ptarmigan Glacier system is close to equilibrium, and the rest are slowly down-wasting and in a state of very gradual terminal retreat, several of which have recently approached equilibrium positions in their terminal zones. The relative locations of these glaciers are noted in Figure 1.

For purposes of this research, a network of 13 synoptic meteorological field stations and ablation sites were established. These range from low-level to high maritime conditions on the coastward flank of the icefield and from high-level to intermediate-level conditions on the continental interior of the Alaska-Canada Boundary Range. Concurrently, the regional research has proceeded on rock glacier development and on frost-processes affecting selected high-elevation features at periglacial sites in zones of both sporadic and permanent permafrost.

An Extension of Previous Studies

In the current program, new procedures have been invoked to extend the scope of previous JIRP investigations along more classical lines. These include the application of IR and Distomat distance measuring equipment in the regional and glacier surveys, the application of wire-embedded strain gages for micro-strain measurements of ice deformation and flow, the use of gravity meter equipment for the detection of "glacier tides", use of portable seismic gear for records of glacier microseisms, the automation of some of our field weather stations with long-running recording units, the establishment of 24-hour watches on avalanche periodicities, the application of electrical resistivity methodology in the delineation of glacier water-tables and sub-surface ice depths in localized frost-mound areas, and so forth.

Of unusual interest is the new trend of névé and glacier fluctuations being manifest by regional cooling since the mid-1960's, a critically important trend now being reflected also in other areas of the Arctic. The significance of this is that coincident with the current downward trend in transient snow-lines (and in late summer the annual seasonal névé-lines) revealed since the 1950's, there has also been a notable inland shift in storm tracks over these years. The result has been an increasing maritimity at the more inland sites on our station transect. The station meteorological records, the test-pit data, crevasse-wall stratigraphy and névé-line information obtained in these past four years have, therefore, been very critical in up-dating these significant assessments, which as part of the on-going JIRP contribution have considerable importance with respect to the overall North American Arctic and sub-Arctic.

The significance of this research also lies in bringing into focus our understanding of fundamental glacio-climatic and sea-ice trends over large areas of the Arctic as well as on a global scale. Because the investigations have focused on snow and ice variations with short-term periodicities which are superimposed on the more secular trends, they especially relate to the climatological cause of these smaller fluctuations which have direct and immediate influence on the terrain.

As the [illegible] the synoptic and regional data collected will be incorporated in future records and thus eventually treated in more detail than can be reported here, attention is given to those more concise and "easily packaged" investigations of a selective nature which are ready to be reported upon. It is to be expected that some of these studies are separate entities worthy of note on their own and because of this that undue emphasis may be placed on some sectors of the total systems than on others. Although there have to be some gaps with respect to the coverage, we are confident that in subsequent reports and in already published as well as anticipated papers in professional journals these gaps will be filled. Conversely, this report fills some of the gaps between the published records of the past and the future.

Each of the following main phases of research is inter-related scientifically and as well has been logistically integrated in the field to permit maximum results at minimum cost. The main categories of research as noted in the table of contents are (1) an overview; (2) glacio-climatology; (3) glacio-hydrology; (4) glaciology, mass balance; (5) surveying and photogrammetric mapping of glacier surfaces and flow; (6) geophysical determination of ice depths and buried bedrock terrain; (7) continuum mechanics of glacier bulges and structural glaciology; (8) avalanche research; (9) periglacial features and processes; (10) patterns of glacier termini variations in the Neoglacial; (11) Pleistocene terrain features, erosional and depositional; (12) major geomorphic elements of the terrain, their structural and lithologic control; and (13) the total systems view in terrain analyses.

The topics are dealt with in order. Research personnel and their affiliations are credited on the first page of each section.

The list of publications and theses resulting from this study are listed in Appendix A.

Part II GLACIO-CLIMATOLOGY

A. SYNOPTIC REGIONAL METEOROLOGY OF THE JUNEAU ICEFIELD*

During each of the past four summers, standard meteorological records have been obtained at 13 stations on the Icefield between Juneau, Alaska and Atlin, British Columbia (Figs. 1 and 2). As well, government operated stations at the Juneau Airport, Alaska and the Whitehorse Airport, Yukon Territory have provided year-around records and useful base station reference for interpretations concerning coastal and inland weather conditions. The data from Juneau City (elevation 75', not the same as Juneau Airport), Annex Creek (sea-level), Eldred Rock (sea-level), Haines and Skagway, Alaska, as well as Teslin, B.C., can be incorporated for refined regional interpretations (v. Miller, 1972). Such a broad regional synoptic treatment, however, is beyond the scope of the material presented in this report.

At most of the field stations, data have been taken between July 1 and September 1, with comparative data also obtained at Camps 17 and 29 from May 31 to September 15 and at the Atlin Station (Camp 30) on a year-around basis since 1968.

Essential field meteorological equipment (including radiometers) have been provided at each station by the Foundation for Glacier and Environmental Research. Automatic recording stations, used at Camps 17A, 17, 10, 16, 8, 26 and 29, were provided by the National Atmospheric Research Center (NCAR) at Boulder, Colorado. This has permitted continuous surface weather and ablation/accumulation records to be obtained at at least 10 field stations from the coast directly across the Juneau Icefield to Atlin Lake. This transect is some 130 miles long and involves quite contrasting climates. These range from extreme maritimity (precipitation 90+ inches/year) at Juneau, Alaska to semi-arid continentality (precipitation of 10+ inches/year) at Atlin, British Columbia.

A brief description of all 15 stations used in this study is tabulated on the following page (also see transect map in Fig. 2 and selected photos in Appendix H. The asterisked stations are fully equipped JIRP weather stations with permanent buildings and other facilities for allied glaciological research activities. Stations 18B and 29B are temporary sites used for short-duration micro-meteorological and hydrological studies during a few weeks each summer.

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* Scientific expertise for this aspect of the program has been provided by Dr. A.H. Thompson, Professor of Meteorology, Texas A and M University; Dr. Edward Little, formerly senior Ice Physicist, Arctic Submarine Laboratory, Naval Underwater Research Center, San Diego; and Dr. Maynard M. Miller, Professor of Geology, Michigan State University.

TABLE I

GLACIO-METEOROLOGICAL STATIONS
USED IN THIS RESEARCH PROGRAM

Site IA	Elev. 150', near sea level, a U.S. Forest Service station at Mendenhall Glacier terminus; climatological conditions low-level, wet maritime
Camp 17A*	Elev. 2500', on coastal flank of Boundary Range yet still relatively low elevation near timberline; wet maritime conditions
Camp 17*	Elev. 4300', on Lemon-Ptarmigan Glacier system; intermediate elevation maritime climate
Camp 16*	Elev. 5000', on intermediate elevation névé of Taku Glacier; sub-maritime interior icefield climatic conditions
Camp 9*	Elev. 5200', on upper Matthes Glacier (North Central Branch, Taku Glacier); higher level, sub-maritime to sub-continental climate
Camp 18*	Elev. 5600', on joint névé of Upper Taku and Vaughan Lewis Glaciers; sub-maritime to sub-continental high level conditions (ice divide and transition zone)
Camp 18B	Elev. 3600', on lower Vaughan Lewis Glacier, at head of large valley with drainage to coast; intermediate elevation sub-maritime conditions
Camp 19*	Elev. 3800', on Watchamacallit Glacier in the Gilkey Canyon near site 18B; intermediate elevation, sub-maritime conditions
Camp 25*	Elev. approx. 7000', on Mt. Nesselrode and the broad névé plateau of the Bucher Glacier, part of which also flows into the Llewellyn Glacier drainage; high-elevation continental and Polar conditions
Camp 26*	Elev. 4700', at intermediate elevation on Llewellyn Glacier, on interior flank of Boundary Range; glacial sub-continental to continental climate
Camp 29*	Elev. 5300', in permafrost zone on Cathedral Glacier Massif, west of Torres Channel, Atlin Lake; continental, semi-arid, sub-Polar conditions
Camp 29B	Elev. 4400', at timberline and in alpine tundra zone near Cathedral Glacier terminus; continental climatic conditions
Camp 30*	Elev. 2300', on Atlin Lake, 130 miles due north of Camp 17, on the inland flank of the Boundary Range. Continental, semi-arid climatic conditions. Surrounding hills exhibit periglacial and sporadic permafrost conditions.

Whitehorse Airport Elev. 2200',Canadian Met. Service

Juneau Airport Elev. 20', National Weather Service

* Fully equipped JIRP weather stations with permanent buildings and other facilities for allied glaciological research activities.

At each station, a senior scientist or graduate student , with several trained younger field assistants, was placed in charge to assist in the basic, three-hourly synoptic weather and glaciological measurements required in this program.

Procedures used by the National Weather Service were invoked in all weather recordings. In the ice mass balance measurements, the procedures recommended by the International Commission on Snow and Ice of the International Hydrological Decade were used. The synoptic data were obtained directly by assigned observers daily at 0700, 1000, 1300, 1600, 1900 and 2200, with the night-time data recorded by instruments.

In 1971, 1972 and 1973, Dr. E. Little, Dr. A.H. Thompson, Dr. A. Jahn and T. Dittrich conducted radiation studies on the nature and cause of ablation and sun-cup generation and degeneration at Camps 10, 26 and 29, under varying cloudy, rainy and clear conditions, to compare with observations made on the maritime snow-pack and firn-pack at Camp 17 in 1969 and 1970 (v. Miller, Dobar and Wendler in Miller, 1972b). Some aspects of this research will be reported in future journal articles. Dr. Pinchak, assisted by W. Lokey and others, continued ice ablation and thermal erosion studies at Camp 18B which allied with the meteorological records. Aspects of this research are presented in Part II of this report.

B. CONTINENTALITY ACROSS THE JUNEAU ICEFIELD IN STORMY AND FAIR WEATHER*

Weather observations made at 10 of the key field stations on the Juneau Icefield (Fig. 1) during the summer of 1971 were used to investigate the variation of weather and continentality along an approximately 200 mile south to north profile extending from Juneau across the Icefield via Atlin, B.C. to Whitehorse, Y.T. (Fig. 2). Selection of a day during fair weather and one during a storm allowed comparison of the gradient of continentality under both types of conditions. The effect showed up on the profile for both situations and was generally more pronounced for the fair weather case at the icefield stations. Most surprisingly, Juneau, nearest the ocean, showed about the same continentality in some measures as the inland station at Whitehorse. Otherwise, the main gradient seemed located on the inland side (or more properly the continental flank) of the divide along the United States-Canadian border.

On the stormy day, Juneau appeared to be the most maritime in characteristics, while the region of maximum gradient appeared to be located at a considerable distance inland from the divide, being north of the northernmost glacier station. Despite onshore flow throughout the storm period, the precipitation amount was quite uniform across the icefield and the northern end of the section. The maximum precipitation was at the southern end of the profile, although a near-record 24-hour rainfall for summer occurred at Whitehorse.

A Synoptic Approach

Much of the work utilizing the meteorological observations made on the Juneau Icefield has tended to be based on climatological methods to examine the data. The correlated glaciological studies have required the data gathering and evaluation techniques of the climatologist. The instrumentation of the various stations proved to be most suitable for such climatological observations. Some limitations arose, however, because some of the observers lacked the training and experience required to take accurate synoptic-type observations, and there were occasional significant problems with the data gathered.

However, the sets of climatological data are made up of the many events of individual days. Further, the scientific and technical workers engaged in the various studies worked under the actual day-to-day weather conditions which may perhaps better be considered as synoptic meteorology. With this appreciation, the conditions of a few days are discussed in some detail. Some emphasis is placed on information related to the gradient of the continentality across the icefield, as investigati of this problem was one of the objectives of the program in the summer of 1971.

The last few days of July, 1971 were characterized by fair weather with the 31st being nearly clear at all stations on the icefield. During the next two days the weather conditions deteriorated as a storm approached from the Gulf of Alaska. Rain and snow became quite general by August 2nd. These three consecutive days provided a good example of the diversity of conditions on the icefield. The storm was not characterized by strong winds. Since a significant number of the summer storms do have strong winds, a few remarks are added concerning such cases.

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* Prepared by Dr. Aylmer H. Thompson, Dept. of Meteorology, Texas A and M University

The Setting and Instrumentation

The series of research camps used in this study is shown in Figure 1. The number of camps and observation sites has changed from year to year, but those used in 1971 for this study are indicated in Table II. Pertinent information about each site is also included in this table.

Instruments at the stations included maximum and minimum thermometers, sling psychrometers, thermographs or hygrothermographs, rain gauges, and recording wind direction and velocity indicators. Four stations included MRI Mechanical Weather Stations. Five stations had Campbell-Stokes sunshine recorders and Belfort mechanical actinographs. Cloud information, visibility and weather were observed visually. Some sites were manned only occassionally, the recording instruments providing records for periods when such sites were not manned. (The instrumentation is also summarized in Table III.)A few data gaps occur due to inability to change records on time during periods of severe weather, but these are not prejudicial to the general synoptic interpretations.

Characteristics of a Clear Summer Day

Southeastern Alaska was dominated by relatively high sea-level pressures during the last few days of July, 1971, with a high center of pressure (1021 mb) about 600 km to the SSW of Juneau at 1200 UT (0500 PDST) on the 31st. At the same time, a low center was present over the eastern Aleutians and a weak, slowly moving front extended south from Kodiak Island. As the lower tropospheric cold air was over British Columbia, the mid-tropospheric flow over Southeastern Alaska was from the SW, with anticyclonic curvature characteristic of a ridge. These and other features may also be seen by studying the weather charts for that date.

The weather associated with the generally anticyclonic conditions was good over the Alaska Panhandle and Northwestern British Columbia, although cloudiness preceding the low was extensive over the remainder of Alaska. The sets of observations on the 31st from all stations are summarized in Table III.

The characteristic weather and clouds for the camps are summarized in the lower middle of the table. Winds were light and variable. All stations reported small amounts of cirrus, seldom more than a tenth of the sky being covered. In addition, small amounts of fog or stratocumulus occurred in the valleys around Juneau, south and west of the Icefield.

With such cloud conditions, precipitation should be essentially nonexistent... as was the case. Camp 17 observers detected a trace during the night of July 30th, robably a little drizzle associated with night fog. All other stations reported essentially 100 per cent of possible sunshine, despite occasional cirrus. Actual hours of sunshine showed some variation between sites due to the presence of different blocking effects from nearby mountains. (The hours of sunshine reported for Juneau are based on the assumption that no mountains are present.) The way in which the sunshine recording sheets indicate such cutoffs is illustrated in Figure 3a a tracing of the actinograph recording for Camp 26 on July 31st. The sun rose about 0600 PDST, and the trace shows the gradual and steady increase in intensity of the solar radiation. The evening portion of the trace appears much different, with the drop starting gradually in mid-afternoon, then plunging suddenly about 1820 PDST as the sun dropped behind Corona

Peak, immediately west of the Camp 26 research station. The minor irregularities, including the sharp drop in radiation about 1045, are due to small amounts of denser cirrus reducing the solar energy reaching the recorder. The Campbell-Stokes record indicates the same features, except that the cirrus are usually not dense enough to indicate their presence on the record.

Comparison of temperatures at the various sites is difficult at best. Elevations range from sea level to 2200 m and exposures are not uniform because of the difficult terrain.* Nonetheless, the reported values provide some interesting comparisons. As would be expected, the highest station, Camp 8, is the coolest and the low inland station at Whitehorse is the warmest. The temperature values seem to divide the stations into groups, with Juneau and Atlin in one group; Camps 10, 17, 18 and 26 in another; and Camps 8 and 9 in a third. Camps 10 and 17 are near an elevation of 1200 m, while Camps 18 and 26 are close to 1600 m. Similarly, Camp 8 is about 640 m higher than Camp 9, and Atlin is about 670 m higher than Juneau. Since the exposures of the instruments at various stations are not that different, the elevation differences should not lead to the above-noted groupings. The continentality effect of Atlin and Camp 26, however, could lead to the observed similarities with Juneau, Camps 10 and 17 respectively. The reason that Camps 8 and 9 have such similar temperatures despite the difference in elevation is not immediately clear.

The values of the range of temperature at each station are worthy of attention. The three lowland stations have the same range, while the icefield stations have a much smaller range. This may be due in part to the locations on nunataks where the coldest air is able readily to drain away with night cooling. The range is greatest at the most interior icefield camps (e.g., C-26); otherwise no systematic variations are obvious for the icefield stations.

Thermograph records for Camp 26 on the 31st show a number of interesting non-periodic temperature increases (often on the order of 10°F.) which last from a few minutes to over an hour. A portion of the hygrothermograph trace is reproduced in Figure 3b. The normal diurnal effect is present, but tends to be suppressed by the sudden warmings, which seem to occur at any hour and are accompanied by a corresponding decrease of the relative humidity. In some cases, as near 1600 PDST, the change is quite sharp and definite, amounting to about ten degrees in as many minutes. In most cases there is no appreciable change of other variables recorded. The only exception may be an increase in the variability of wind direction but without an effect on the speed. There may also be some tendency for the air flow to be more from the east than any other direction. Since such a flow would be both upslope and off the glacier, it is difficult to see why warming would occur. The same phenomenon occurred on many other days at this camp, usually on days of fair weather, but apparently did not occur at other camps. No logical explanation is apparent now, but the phenomenon should be looked for during future summers.

On this date, July 31st, Whitehorse experienced the highest temperature of the summer and a record high temperature for that date.

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* The original data were recorded in the Fahrenheit scale and have not been converted to Celsius. Such conversion either implies unwarranted accuracy, if given to the nearest tenth of a degree, or excess inaccuracy if given only to the nearest whole degree Celsius.

Characteristics of a Cloudy Summer Day

During the following two days, the pressures over the icefield area dropped gradually as the low from west of the Alaskan Peninsula moved eastward to south central Alaska and the Kodiak Island area. At the same time, the highs retreated southwestward deeper into the Pacific and southeastward into the northwest United States, while the low, north of the Alaskan Arctic coast, migrated southeastward into southern Yukon Territory. The weak gradients and intensities of these systems are fairly typical of summer systems. *

The discussion in the remainder of this section will consider conditions on August 2nd, to contrast with the discussion of the clear day on July 31st. The observations for the 2nd are summarized in Table IV. Direct comparisons may be made with the data for July 31st in Table III.

Appreciable cloudiness began to develop and ceilings to drop on the icefield near noon on August 1st. Rain (or snow at higher elevations) began shortly before midnight over the south and west parts of the icefield, spreading over the entire icefield and on to Atlin and Whitehorse early on the 2nd. Precipitation continued through most of the 2nd and much of the 3rd.

The elevation at which the snow changed to rain was about 1600 m. on the maritime side of the range, and probably not much higher on the continental side. Camp 8 reported nothing but snow, while Camp 18 reported rain mixed with and changing to snow as the day progressed. The action of the precipitation recorder at Camp 16 suggests that either the precipitation fell as rain or the snow melted almost as rapidly as it fell. At all other lower camps, only rain was reported.

The amount of precipitation on 2 August was remarkably uniform. Except for Juneau, Camps 17, 18 and 8, the stations measuring this element reported close to 0.9 inches. (The measurement at Camp 8 is suspect. There was much drifting snow and the observers were not skilled in melting the snow to obtain a measure of the water equivalent.) This uniformity across the range suggests a lack of any appreciable rain shadow effect, at least for this summer storm. The comparatively larger amount of precipitation at Juneau and the still larger amount at Camp 17 suggest that there is an appreciable upslope effect in that vicinity. The terrain near these two stations is probably as favorable for such an effect as any location in the area. One may hypothesize that the general convergence of the low pressure system resulted in rather uniform precipitation over the area and that the maximum upslope action near Juneau and Camp 17 superimposed an added amount of precipitation at these particular stations.

With the moisture and convergence of the low-pressure system, clouds were thick and low. All icefield camps reported fog, except for Camps 10 and 26 and the stations off the icefield which reported low stratus, stratocumulus and nimbostratus, with ceilings below 700 m. Thus the clouds were enveloping the higher mountains. The comparatively high ceiling at Camp 26 at first appears to be a lee side or continental effect, but the much lower value at Atlin is not consistent with such an interpretation. It is difficult to ascribe this difference to an error in estimating the height of the cloud base above one of the stations, as the temperature and moisture values are consistent with the observed heights. Further, a note made by the observer at Camp 26 indicates that he could see snow collecting on the rock near the tops of the nearby peaks The problem remains that the nearby peaks are only about 300 to 400 m above camp and the tops

* A comprehensive review of the regional and hemispheric relationships of such tropospheric systems and shifts in their related ridges and troughs, as they relate to secular glacio-climatic trends on the icefield, has been completed by Jones (1975).

were almost certainly obscured by clouds. Most certainly, the lee effect is not obvious.

As a result of the extensive and persistent cloud cover, no sunshine was recorded at any stations. The short-wave energy received at the surface was correspondingly very low.

Wind directions tended to have a southerly component at all stations except Juneau, although the actual direction at any observation point seemed to be determined by local terrain. This was particularly noteworthy at Camp 18, which is sheltered from both northerly and southerly winds and the dominant directions are either up or down the Gilkey Glacier Trench (west of Camp 18). The channel effect at Juneau is obvious.

Thermometric measurements from the various stations were not as regular as the nature of the clouds and precipitation might suggest. First it should be noted that the values in the table refer to the 24 hours beginning at midnight and were determined from thermograph traces or hourly observations. (An exception was Camp 17, for which no thermograph trace is available. The temperature at 2200 PDST on August 1st is used as the maximum.) The expected drop in temperature with elevation is present, although not completely consistent. The very high maximum temperature at Whitehorse results in that station having the highest average temperature. Further, Camps 17, 10 and 26 have the same values for high and low temperatures, despite the range of elevation of about 300 m. (Camp 18 would fit this group also except for a lower minimum.) This is a characteristic similar to that noted for the 31st. Atlin was a little cooler than Juneau. Much of the inconsistency may again be ascribed to continentality. Worth mention is the 24-hour cooling evident for all stations. This drop amounted to 5 and 6 degrees for Camp 26 and Juneau, but was larger at all other camps, reaching 13 degrees at Camps 8 and 18, and 16 degrees at Whitehorse.

The temperature and humidity traces for much of August 2nd at Camp 26 are shown in Figure 3c. Noteworthy is the smoothness of the trace, compared with a clear day (see Fig. 3b). Even on the cloudy day, however, there are numerous fluctuations of about a degree in amplitude and a few minutes duration. Corresponding fluctuations in relative humidity also occur. Note that the humidity fluctuations of only a few minutes duration are of about the same amplitude on both the clear and the cloudy day, whereas the corresponding fluctuations in temperature are much less on the cloudy day than on the clear day. These short-period fluctuations indicate that the air is not uniform even in a very small-magnitude sense,but rather is a turbulent, non-homogeneous mixture. These are typical characteristics of the atmosphere. The traces for the other stations are similar, also showing nonhomogeneity (Thompson, 1974).

The effect of continentality is again present, although not so obvious as on the clear day. The change seems to be confined primarily to the region between the divide (Camp 8) and Camp 26. The effect is not strong, and the dominance of the convergent effects of the low pressure system and storm, which is really not an intense disturbance even for the summer season, is probably most significant when compared not only to the continentality but also to local effects.

Strong Winds with Storminess

Of particular interest to the personnel working on the Icefield are the strong winds occurring at some stations during some of the storms. As would be expected, these winds are stronger at the higher stations. Camp 8, a short distance below the summit of Mt. Moore, nearly always reports the highest velocity winds. Camps 9, 16, 17 and 18 also tend to experience strong winds. Camp 26, while higher than two of these stations, is in a slightly sheltered location and probably does not receive the full effect of the air motion. The actual winds reported at the various stations depend on local terrain effects as well as on the direction and strength of the gradient of pressure over the Icefield.

Some of these points are illustrated in the storm of August 2nd and thus in the wind data of Table IV. These data represent the prevailing wind direction and average speed for the individual stations. The maximum sustained speed at most stations was roughly twice the average value. These data may not include the maximum gusts, since the limitations of the instruments precluded reliable estimates of gusts. There is some indication that this condition pertained generally throughout the summer on the Icefield, although the factor of 2 relating average to maximum speed should not be relied on too exactly. On August 2nd, Camp 8 reported the strongest wind, although Camp 16 was not much weaker. The strongest winds were from the south, from which direction Camp 18 is sheltered due to its location at the head of the east-west oriented Gilkey Trench (note the weak winds at Camp 18 on August 2nd, as well as the existence of two prevailing directions). The Northeast wind at Juneau has been mentioned earlier.

Stronger winds occurred during the summer, in one case approaching an estimated speed of 60 mph at Camp 8. On that date (19 August), the flow was more from the east and Camp 18 also reported strong easterly winds with gusts occasionally exceeding 40 mph. The usual cause of such winds is similar to the case of August 2nd...namely a low crossing the Gulf of Alaska and approaching the coast of Southeast Alaska. The strength and direction of the winds is closely related to the magnitude of the pressure gradient and to the cyclonic track of the pressure system as it approaches and crosses the coast.

Some Conclusions

Summer weather conditions across the Juneau Icefield area are characterized by reasonable consistency and an appreciable tendency for uniform conditions at a given time. Vertical motions due to the dynamics of the meteorological flow pattern appear to play a greater role than the terrain features and the effect of continentality in determining the actual weather and its spatial variations. The changes of the meteorological elements across the Icefield caused by terrain and the continentality effects show up rather well in the statistical summaries (and, in fact, are quite obvious to the scientist travelling across the Icefield). They can become masked and quite subtle, however, in the individual cases such as those considered here.

TABLE II

OBSERVING SITE DESCRIPTION AND INSTRUMENTATION

	Juneau	Camp 17	Camp 16	Camp 10	Camp 9	Camp 18	Camp 8	Camp 26	Atlin	Whitehorse
Elevation (meters)	4	1275	1540	1177	1550	1710	2195	1585	670	701
Physical setting	Airport	Arété	Small Nunatak	SW Slope of Mt.	Small Nunatak	Rock Cleaver	West Slope of Mt.	East Slope of Mt.	Lake Shore	Airport
Dates of occupancy (1971)	All year	5 May to 10 Sep.	OCNL	14 Jun. to 9 Sep.	OCNL	23 Jun. to 31 Aug.	21 Jul. to 31 Aug.	21 Jul. to 31 Aug.	5 Jun. to 31 Dec.	All year
Wind instrument	U.S. Standard	Indi-cating	MRI	MRI	MRI	Indi-cating	Indi-cating	MRI	Indi-cating	Canada Standard
Sunshine recorder	Eppley	C-S	-	C-S Belfort	-	-	C-S Belfort	C-S Belfort	C-S Belfort	Eppley

Key to instrumentation:

"Indicating" -- An anemometer and wind vane combination with some form of direct reading dials. Several brands were in use.

"MRI" -- The automatic weather station of Meteorology Research, Inc.

"Eppley" -- The standard Eppley pyrheliometer or a modification thereof.

"C-S" -- The standard Campbell-Stokes sunshine recorder

"Belfort" -- The mechanical pyrheliograph manufactured by Belfort Instrument Co.

TABLE III

TYPE FAIR WEATHER CONDITIONS IN THE JUNEAU ICEFIELD AREA ON 31 JULY 1971

	Juneau	Camp 17	Camp 16	Camp 10	Camp 9	Camp 18	Camp 8	Camp 26	Atlin	Whitehorse
Maximum temperature (°F)	78	61	-	63	55	60	55	65	78	86
Minimum temperature (°F)	48	48	-	48	43	50	42	45	48	56
Average temperature (°F)	63	5[illegible]	-	56	49	55	48	55	63	71
Range of temperature (°F)	30	13	-	15	12	10	13	20	30	30
24-hr temperature change	-1	-	-	+1	+1	-3	-8	+6	-6	+2
Relative humidity (per cent at 1300 PDST)	68	88	-	76	95	81	62	44	62	30
Representative weather	Fair	SCTD Clouds	-	Few Clouds	-	Few Clouds	Few Clouds	Few Clouds	Few Clouds	Few Clouds
Representative clouds	$1/_8$ Ci Sc	$1/_8$ Ci Sc	-	$1/_8$ Ci	-	$1/_8$ Ci	$1/_8$ Ci	$1/_8$ Ci	$1/_8$ Ci	UNK
Average heights of clouds (hundreds of feet)	250	250 5	-	250	-	300	250	300	250	UNK
Prevailing wind velocity (dir / speed in mph)	NW/6	Calm	SW/9	VRBL/4	SE/7	VRBL/2	S/12 (est)	WSW/15	S/8	SW/11
Hours of sunshine	16.6	15.8	-	14.8	-	-	14.8	12.3	15.5	14.8
Percent of possible sun	100	95	-	100	-	-	100	97	98	89
Precipitation (inches)	0	T	0	0	-	0	0	0	0	0

TABLE IV

TYPE STORMY WEATHER CONDITIONS IN THE JUNEAU ICEFIELD AREA ON 2 AUGUST 1971

	Juneau	Camp 17	Camp 16	Camp 10	Camp 9	Camp 18	Camp 8	Camp 26	Atlin	Whitehorse
Maximum temperature (°F)	59	51	-	51	45	50	44	51	55	71
Minimum temperature (°F)	52	41	-	41	31	33	29	41	45	48
Average temperature (°F)	56	46	-	46	38	41	36	46	50	60
Range of temperature (°F)	7	10	-	10	14	17	15	10	10	23
24-hr temperature change	-6	-8	-	-9	-12	-13	-13	-5	-13	-16
Relative humidity (per cent at 1300 PDST)	93	100	-	93	100	100	100	70	100	98
Representative weather	Rain	Rain	-	Rain	-	Rain Snow	Snow	Rain	Rain	Rain
Representative clouds	Ns, Sc	Fog	-	Stratus	-	Fog	Fog	Not given	Stratus	Stratus
Average heights of clouds (hundreds of feet)	25	Fog	-	10	-	Fog	Fog	20	10	UNK
Prevailing wind velocity (dir / speed in mph)	NE/6	SW/10	SW/16	ESE/4	WSW/7	E/4 WSW/4	S/21	SSE/13	SSE/10	SSE/7
Hours of sunshine	0	0	-	0	-	-	0	0	0	0
Per cent of possible sun	0	0	-	0	-	-	0	0	0	0
Precipitation (inches)	1.63	2.00	0.85	0.82	-	1.43	0.47	0.90	0.91	0.95

C. COMPARISON OF PERCENT POSSIBLE SUNSHINE AT FIELD STATIONS ON THE JUNEAU ICEFIELD *

In this section analysis is made of data obtained from Campbell-Stokes sunshine recorders positioned at seven field stations along a north-south transect across the Juneau Icefield. Particular study of these data was made between July 22nd and August 31st, 1971, to determine the percentage of possible "bright sunshine" experienced at each of the reporting stations. The locations of the observing stations are indicated on the location maps in Figures 1 and 2. This technique has been applied in earlier JRIP research, especially in comparison of winter vs. summer duration of sunshine at Juneau and Camp 10 (Miller, 1953).

In this study, the time at which the sun struck or was shadowed from the recorder was determined for each station on each day the station experienced clear sky at either sunrise or sunset. From these data it is possible to plot the beginning and ending of possible sunshine for each station. In addition, the total possible sunshine is computed for the average station latitude of 59° N., thus giving a reference from which each station's possible sunshine can be computed. This permits us to account for the effects of local topographic shading, which is particularly significant at Camps 8, 10 and 26 (Fig. 1).

The percent possible sunshine can be determined in one of two ways when accounting for the topography. (1) It can be calculated as a percent of the time the sun is above the local orographic obstacles as seen from the observing point. This would give different daily possible and total possible hours of sunshine at the various stations. (2) The actual hours of sunshine can be extrapolated to astronomical sunrise and sunset times, accounting also for the potential effect of observed cloudiness during the time the field site is shielded from possible direct sunshine by orographic barriers. This would give the same daily possible and total possible hours at each station, except for a small latitude effect (which is neglected here). As the latter procedure is most useful when comparing stations in a given locality having different shading effects, this method is employed.

Both actual percent possible sunshine and the corrected total possible percentages, allowing for the shadowing effects of local topography, are presented in Table IV. The data for Juneau, Alaska and Whitehorse, Y.T., are included to provide control points on the south-north transect (Fig. 2). It should be noted that the Whitehorse data are for duration of "bright sunshine" , and are not corrected for local shadowing effects because there are no nearby obstructions which would introduce shading at this time of year except within a few minutes of sunrise or sunset. The Juneau data are already corrected for shading effects.

The period between astronomical sunrise and astronomical sunset varied from 17.4 hours on 22 July to 14.3 hours on 30 August, 1971. Over this period, there were 641 total hours of possible sunshine at an unshielded location (Table V).

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* Prepared by Stephen C. Andrews, Dept. of Geography, University of North Dakota

The evaluations of the percent of possible sunshine at the several stations along the transect are prepared in tabular form. An example, sufficient to illustrate the application, is presented in Table VI for Camp 8. It should be noted that on many days with much sunshine it was nearly clear at sunrise and/or sunset so the maximum correction was applied. On days with only small amounts of sunshine, however, conditions were such that even if no orographic barriers had been present there were sufficient clouds near sunrise and/or sunset that no correction needed to be applied. The results of the evaluations of percent of possible sunshine for all stations along the transect are presented in Table V and the conclusions discussed below.

Juneau Airport (20 feet)

Juneau was found to have the lowest percentage of possible sunshine (22.6%). This is due to the high incidence of fog and rain, resulting from the strong maritime influence at that station.

Camp 17 (4200 feet)

Since Camp 17 has no appreciable shading, correction was not applied to these data. Although its percentage of "bright sunshine" (24.4%) was higher than Juneau's, it has the lowest percentage on the icefield. This low percentage is a reflection of the extreme maritime influence at this station.

Camp 10 (4000 feet)

Camp 10 had a low amount of possible sunshine (28.4%) before correcting for the sunrise shading by the mountain to the east, Taku B. After correction, the value rose to 32.7%. It was assumed that Camp 10's sunset occurred at astronomical sunset, even though this results in a small error. (Astronomical sunrise is determined by subtracting the hours of daylight from the observed time of sunset.) Camp 10's lack of sunshine appears to be due to its elevation and again its position within an orographic maritime system (Thompson, 1974).

Camp 8 (7200 feet)

At Camp 8 there was also a low percentage of possible sunshine, presumably due to its position on the crestal plateau of the Juneau Icefield. Camp 8 had an uncorrected percentage of possible sunshine of 24.4%. Thus, Camps 8 and 10 are shown to be essentially similar. Both camps are under the climatological influence of orographic lifting. Since Camp 8 also experiences sunset near astronomical sunset, the calculation to account for the difference between astronomical sunrise and actual sunrise was made in the same manner as for Camp 10.

Camp 26 (4700 feet)

For the Camp 26 data it was necessary to correct for the shading effect of nearby Corona Peak, directly west of camp. As suspected, Camp 26 has the maximum percent of possible "bright sunshine" of any camp on the icefield. Although Camp 26 had 40.9% uncorrected possible sunshine, when corrected for the shading effects of Corona Peak at sunset, this camp had a total of 57.0%.

Camp 30 (Atlin, 2180 feet)

Of the continental stations on the transect, the Atlin meteorological station had the lowest possible converted percentage of "bright sunshine". Atlin is not shaded to any degree and therefore no corrections are introduced into the data. The percentage of "bright sunshine" for the Atlin station was 49.4%, compared to 57.0% for Camp 26 and 52.8% for Whitehorse.

Whitehorse Airport, Y.T. (2300 feet)

Whitehorse Airport, at the northern end of the transect, lies on a high flat-topped bluff overlooking the Yukon River. In this position shading effects are negligible. Thus local sunrise and sunset times are assumed to have been near astronomical sunrise and sunset. On this basis, Whitehorse received 52.8% possible "bright sunshine".

Summary

The presence of a "sunline", i.e. a boundary between areas of low possible "bright sunshine" and areas of high possible "bright sunshine" is indicated in that section of the transect between Camp 8 and Camp 26 (i.e., close to the orographical divide a few miles north of the Alaskan-British Columbia border).*

Although there are two climatic regimes represented by these seven stations (1) maritime (Juneau, Camp 17, Camp 10, Camp 8) and (2) continental (Camp 26, Camp 30 at Atlin, and Whitehorse), with probably orographic lifting effects influencing the conditions at Camps 17, 10 and 8, there does not appear to be any out-of-phase relationship between these two basic regime zones. After a spot check of several days when new fronts appeared to be moving into the area, it was found that the time between the start of pressure and precipitation effects at Juneau and the start of the same effects on the opposide side of the Icefield was measured in terms of hours instead of days. It is probable that out-of-phase relationships do exist, but it is now clear that a much smaller scale is needed to measure them.

From the sunshine data it is concluded that similar conditions exist almost simultaneously across the Icefield when major frontal movements are involved. In weaker systems, activity is dissipated before crossing the crestal plateau of the Icefield and reaching the continental sector (revealed by these interpretations to lie at and northwards into the Atlin/Teslin/Whitehorse area (Fig. 2).

For future investigations, an additional field meteorological station should be installed between Camp 8 and Camp 26 to enable a better determination of the transition effects that occur in that area. This will allow more accurate location of the "sunline". The establishment of Camp 25 at 7100 feet elevation on the Nesselrode Plateau in 1973-74 and subsequent data from that site will facilitate this refinement in the study.

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* In 1973 and 1974, a new research site (Camp 25) was developed on the 7000-foot névé of the upper Bucher Glacier near Mt. Nesselrode (Fig. 2). Summer-time weather records at this site suggest that the sunline lies close to the summit of this 8100-foot peak on the international border.

TABLE V

COMPARISON OF SUNSHINE RECORDS AT KEY STATIONS

22 JULY TO 30 AUGUST, 1971

STATION	HRS. SUNSHINE	/ CORRECTED	PERCENT POSSIBLE	/ CORRECTED
JUNEAU	145.1	145.1	22.6	22.6
CAMP 17	156.5	156.5	24.4	24.4
CAMP 10	182.1	209.9	28.4	32.7
CAMP 8	156.2	191.8	24.4	29.9
CAMP 26	262.3	365.3	40.9	57.0
ATLIN	316.8	316.8	49.4	49.4
WHITEHORSE	338.6	338.6	52.8	52.8
MEAN	222.5	246.3	34.7	38.4
MARITIME MEAN*	160.0	175.8	25.0	27.4
CONTINENTAL MEAN **	305.9	340.2	47.7	53.1

* Maritime stations: Juneau, Camp 17, Camp 10, Camp 8
** Continental stations: Camp 26, Atlin, Whitehorse

TABLE VI

SAMPLE DURATION OF SUNSHINE RECORD, CAMP 8

DATE	HRS. SUNSHINE	/ CORRECTED	PERCENT POSSIBLE	/ CORRECTED
7/22/71	0.0	0.0	0.0	0.0
23	9.0	9.0	51.7	51.7
24	3.7	3.7	21.0	21.0
25	0.0	0.0	0.0	0.0
26	0.0	0.0	0.0	0.0
27	7.8	11.8	45.8	69.5
28	0.0	0.0	0.0	0.0
29	4.0	4.0	23.5	23.5
30	12.5	16.0	74.9	95.8
31	12.8	15.3	76.6	91.6
8/01/71	9.0	12.5	54.0	76.0
02	0.0	0.0	0.0	0.0
03	0.2	0.2	1.2	1.2
04	2.5	2.5	15.2	15.2
05	3.7	3.7	22.7	22.7
06	0.0	0.0	0.0	0.0
07	0.0	0.0	0.0	0.0
08	9.0	12.3	56.4	77.0
09	12.9	15.9	80.7	99.5
10	11.3	14.3	70.6	90.5
11	12.5	15.5	79.3	98.4
12	12.3	15.3	77.8	97.7
13	0.0	0.0	0.0	0.0
14	0.0	0.0	0.0	0.0
15	0.0	0.0	0.0	0.0
16	0.3	0.3	1.9	1.9
17	0.0	0.0	0.0	0.0
18	0.0	0.0	0.0	0.0
19	0.0	0.0	0.0	0.0
20	1.0	1.0	6.7	6.7
21	0.0	0.0	0.0	0.0
22	0.0	0.0	0.0	0.0
23	0.0	0.0	0.0	0.0
24	0.0	0.0	0.0	0.0
25	0.0	0.0	0.0	0.0
26	0.0	0.0	0.0	0.0
27	12.0	14.3	82.9	98.5
28	4.8	7.0	32.7	46.5
29	4.9	4.9	34.6	34.6
30	10.0	12.3	69.5	84.5
TOTALS	156.2	191.8	24.4	29.9
POSSIBLE	641.0	641.0		

Editor's Note:

It is of glaciological significance that the "sunline" zone concept and the references to maritimity and continentality (noted by Andrews and in the preceding study by Thompson) strongly condition the observed differences in terrain characteristics of the nèvè and ice surfaces to the north and south of this transition zone.

For example, in the Camp 17 to Camp 9 sector, by mid-summer there is a pronounced development of suncups, especially following a number of clear days. After a few days of storm, wind and rain, however, these suncups become subdued and level out, largely by melting processes associated with wind convection and rain... especially when ambient temperatures are above 40°F. In the Camp 8 to Camp 25 plateau sector (5500 to 7500 feet, 1700 to 2300 m), suncup development is essentially nil, but the process is primarily melting even though the nèvè remains quite smooth.

North of the "sunline", on the crestal nèvè, however, a sharp decrease in ablation by melting is observed, coincident with notable increases in evaporation features due to the relatively greater insolation. In fact, at Camp 26 and northward along the Llewellyn and Hoboe Glacier surfaces, radiation weathering features (evaporation spicules) are commonly seen by mid-summer. This is also a pronounced reflection of the decreased relative humidity associated with the increased duration of sunshine reported in this study.

These changes in ice terrain characteristics affect the trafficability of the glacial surfaces with respect to cross-country ski travel and the use of tracked oversnow vehicles.

D. SUMMER TEMPERATURE PATTERNS OF AN INLAND CLIMATOLOGICAL TRANSECT ACROSS THE JUNEAU ICEFIELD *

Maximum and minimum daily temperatures from meteorological records obtained at field stations during the 1971 summer season are analyzed in order to determine some patterns of temperatures and their variations on the Juneau Icefield.

Field Setting

As the Juneau Icefield extends well inland across the high interior of the Boundary Range in the southeast coastal section of Alaska and northwest British Columbia, it consists primarily of high névés, interspersed with barren ice-carved nunataks and nunatak groups separating deep glacier-filled valleys. On the seaward side of the range, to the west and south, some of the glaciers, such as the Taku Glacier, reach sea level, while the termini of many others extend to near sea level in deep glacial valleys and trenches. On the inland (Canadian) side of the range (because of the Taku and Liard Plateaux), they must terminate at higher elevations...i.e., 2400 to 3000 feet (730-909 m). The interior accumulation areas crest at the boundary (water/ice) divide at 6000 to 7000 feet (1818-2120 m), with the elevation of Atlin Lake beyond the northern end of the present ice boundaries being slightly less than 2200 feet (660 m) above mean sea level.

As we have seen, the stations involved in this study lie in an approximate line from SSW to NNE across the icefield. (The key stations are tabulated in Table I.) The lower southwest slopes of the periphery of the icefield near Juneau display a maritime rain forest climate, with precipitation of about 120 inches per year (90 inches at Juneau). The southernmost station, Camp 17, is located on a high rock ridge (arête) between the heads of two valley glaciers, in a high alpine environment at 4200 feet (1270 m), well above the mean névé-line (which lies here at about 3500 feet, 1060 m).

Camp 10 is at 4000 feet (1310 m) elevation in the south interior of the icefield, but as we have seen in the preceding sections is still under the influence of the maritime zone. It is located on a low bedrock shoulder of a nunatak, some 300 feet (100 m) above the névé surface of the Taku Glacier and 18 miles up from its terminus. This surface is also at present well above the mean névé-line (which here lies at about 3000 feet, 909 m). The climate at Camp 8, at 7200 feet (2180 m) elevation, could be termed nival or high arctic. Near the crest of the icefield, this station lies on the western slope of a high nunatak mountain about one mile south of the Canadian border and overlooks the orographic divide at the International Boundary.

Camp 26 is located farther north, some 10 miles beyond the crest and well down the Llewellyn Glacier into British Columbia at the sheltered base of another nunatak mountain. At 4700 feet (1430 m), it is near and sometimes below the inland névé-line in a setting of frost-shattered rock and alpine summer flowers.

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* Prepared by Vernon K Jones, Dept. of Geology, Michigan State University

Camp 30, at 2200 feet elevation (660 m), is located on the shores of Atlin Lake in the semi-arid continental interior where precipitation amounts only to some 12 to 13 inches annually. It is the only field station included in the study which is not on the icefield proper, with the exception of Camp 29 (elevation 5300 ft., 1660 m) which being on a separate and isolated massif in the Atlin area is not considered in this transect comparison.

Observing Procedures

Maximum and minimum temperatures were obtained from two sources. Standard maximum and minimum thermometers were read at 0700 and 2200 by the station observers. Weekly thermograph traces were checked for daily maxima and overnight minima. Previous overnight minima prior to 0700 were used even when a lower temperature was recorded in the evening on a given date. The purpose of this was to standardize thermograph records with maximum-minimum thermometer readings for a given date. Most stations were equipped with fairly new hygrothermographs.

Available data were plotted separately for each station for purposes of analysis. Observed daily extremes from both instruments for a given station and date were compared. Exceptions to this were at Camp 17, which did not have thermograph records until August 1, and Camp 26 which, due to its isolated location, required several days to obtain a replacement for a broken maximum thermometer in mid-season.

Adjusted daily extremes for each station were developed from the two data sources. As the two sources did not always support each other, some careful judgements in the analyses were necessary. In the case of questionable differences, the adjustment tended toward the mean of the two sources. Obviously spurious readings were either heavily discounted or discarded. For example, at Camp 26 occasional suspiciously low readings on the minimum thermometer, when unconfirmed by the thermograph records, were found to be related to overnight wind speeds of 10 mph or greater. Therefore it was assumed that improper positioning of the minimum thermometer, combined with wind-induced vibration of the instrument shelter, had shaken the rider down.

Distinct peaks, troughs, or changes in temperature were generally confirmed by similar changes in both instruments. Occasional gross discrepancies in the two sources were reconciled, based on clues as to any possible causative factors, such as the situation previously mentioned at Camp 26.

Because the data were sometimes obtained by student trainess, they at times lack the precision of those taken by professionally trained meteorologist. The majority of records obtained by both types of instruments, however, are supportive of the observers' records.

Adjusted daily maxima for all stations were plotted on a single graph, allowing synoptic comparison of all stations. The same was done for adjusted daily minima. Daily means were computed for the 41-day period from July 22 through August 31, 1971 when all stations were continuously occupied. These include mean daily maxima, mean daily minima, and a daily mean of the mean maxima and minima. Also computed was the maximum-minimum range of the daily station means. These are shown in Table VII. Included are absolute maxima and minima for each station during the 41-day period of analysis.

Pertinent Climatological Aspects

The Juneau Icefield provides an interesting field laboratory for climatological study...one which has in previous pages been shown as a mountain climate superimposed on a transition from a strongly maritime to a definitely continental climate.

A maritime climate is characterized by a predominating influence of the ocean (in this case accented by the warm Kuroshio Current). The result is high humidity, extensive cloudiness and frequent heavy rainfall interspersed with prolonged periods of drizzle; small diurnal and seasonal temperature ranges; and annual temperature extremes retarded until one to two months after the solstices.

A continental climate has large annual, seasonal,day-to-day, and day-night ranges in temperature; low relative humidity; and a lower occurrence of cloud cover, associated with low to moderate rainfall. Annual temperature extremes tend to occur soon after the solstices.

Superimposed on these climatic characteristics are the characteristics of a mountain climate, with temperatures inversely related to elevation and with definite orographic effects...much of which tends to upset normal lapse rate changes with elevation. Greater daytime insolation and nocturnal radiation may also be expected at higher elevations. In addition, this icefield area ...situated as it is in the Boundary Range... is the interface region between the continental Polar high pressure air mass and the maritime low pressure cellular circulation of the North Pacific. In this region long-term cyclic climatic changes, recorded by glacial fluctuations, have been correlated with changes in the relationship of these air masses and with patterns of solar disturbancs (Miller, 1956, 1963, 1972b, 1973). Dr. Thompson and Mr. Andrews have provided detailed analysis of the continentality gradient in this study area in previous sections.

The Temperature Pattern

As would be expected, Camp 17, on the southwest maritime flank of the icefield shows the lowest mean range of daily mean temperatures...6.7°F. Camp 30 (Atlin), in the continental interior, shows the greatest range...21.0° , followed by Camp 26 on the continental interior slope...13.0°F. Camp 8, near the crest of the range but on the seaward side, shows a range of 9.2° F. These figures are consistent with the gradient of continentality which has been described from maritime to continental climate across the climatic divide.

Andrews has confirmed the continentality gradient in terms of solar radiation (duration of sunshine) received at the earth's surface along this transect. The regional "sun line" which he has described corresponds to the glacial divide, slightly inland from the apparent orographic crest of the Boundary Range and in a clear-cut fashion separates the cloudier maritime flank from the sunnier continental flank.

Camp 10, however, located in the southern maritime interior of the icefield, departs noticeably from this gradient with a mean daily range of mean temperatures of 11.2° F. This value is well above that for Camps 8 and 17, which in the

geographical sense it lies between (Fig. I). Both Camps 10 and 17 consistently experience weather conditions characteristic of a maritime climate, with a relatively high proportion of cloudy, in-cloud, rain, and white-out conditions. This precludes an explanation of greater insolation and nocturnal radiation for the large mean day-night temperature range at Camp 10. Because its position is at the end of a rock shoulder some 300 feet (100 m) above the glacier surface this may provide excellent drainage of denser cold air on the névé surface well below the station level which should prevent minimum temperatures from being unusually low. The location of the instrument shelter on a barren bedrock surface could also allow greater solar heating of the immediate area. But as Table VII indicates, variation from mean maxima and minima at Camp 17 (the most closely comparable station) is almost evenly divided between a higher maximum and a lower minimum. Location of instruments at Camp 17, on a narrow bedrock ridge, provides local conditions similar to those at Camp 10.

In this study a consistent pattern of elevation effects on mean temperatures has been shown, although the pattern does not appear to be linear (Fig. II). Using Camp 8 as the low-temperature end of the scale, daily (24-hour) means are lower at Camps 10 and 17 than at Camps 8, 26 and 30, in proportion to elevation. This may reflect the stabilizing effect of maritime influences on the daily range, although it is not consistent with the unexpectedly high mean daily range at Camp 10.

Special Conditions

It is interesting to note that record or near-record maxima and minima for the study period, on the morning and during the day of August 1st, were followed by secondary record low maxima and minima on the night of August 1st and during the day of August 2nd, during a storm. These had rebounded to near or above seasonal averages by the day and night of August 3rd. Also seasonal record low maxima on August 24-25 and record low minima on August 26 and 27 were followed by secondary record high maxima on August 30th.

In the first case above, the temperature drop accompanied heavy precipitation at all stations. In the second, the temperature drop followed maximum precipitation (or cloudiness at the continental stations) by one to three days, with record lows for the study period accompanying the first CAVU conditions on August 27th. The secondary seasonal maximum for the study period, on August 30th, was associated with conditions of cloudiness or broken cloudiness, with clear weather part of the day at some stations. (Dr. Thompson's previous report has given specific weather details for July 31st and August 2nd.)

Conclusions

Although not all of the data can be given here to support the following conclusions, these data are available in the Foundation for Glacier and Environmental Research's files. The general conclusions are as follows:

Based on detailed analysis of the data, the daily ranges and time-related patterns of temperatures at individual stations indicate the following:

1. Daily ranges of temperature were maximum during clear weather and minimum during poor weather.

2. Temperature minima tended to follow maxima on a day-to-day basis. However, higher maxima from solar heating on clear days tended to be rather obvious. This was partly balanced by nocturnal radiation cooling, resulting in somewhat comparable minima during clear versus cloudy weather. This effect is masked to a large extent, however, by the inherent temperature characteristic of the air mass present at any given time.

3. Temperature minima did not fluctuate as greatly as did maxima.

4. Day-night ranges of temperature were considerably greater in the continental interior than on the more maritime seaward slope of the divide.

5. Temperature variations on a daily interval do not appear to indicate a sequential change of temperature across the Icefield.

6. Differential retardation of annual temperature extremes after the solstice, due to maritime heat-sink effect, does not appear to pertain across this transect.

7. The mean day-night temperature range for Camp 10 departs from the otherwise consistent gradient of continentality.

8. The effect of elevation on mean daily temperatures shows a consistent pattern.

9. Daily temperature variations are for the most part parallel at all stations, suggesting that competent records obtained even over short periods at a single station can have regional significance.

New Inputs from Satellite Imagery

Note is made here of the magnificent aid satellite imagery provides in the regional climatological assessment. As of a few months ago, daily photographs of the North Pacific area are available via the National Weather Service's forecasting offices in Juneau and Anchorage. An illustrative example is given in Figure 4, showing the counter-clockwise swirl of the North Pacific cyclonic low over Southwestern Alaska, with clear high-pressure anticyclonic air dominating the Alaskan Panhandle (photo date , April 24, 1975, 1253 p.m.; NOAA-4 from 1500 km altitude). The value of this input is exponentially increased by the acquisition of simultaneous ground data from key weather stations in Alaska. In our continuing research, considerable use will be made of this combined information.

Of further interest with respect to Figure 4 is that the interaction zone between the opposing pressure systems typefied here is centered over the Alaskan Panhandle from September to December. This is the season when heaviest snow accumulation takes place on the Juneau Icefield. Pertinent aspects of this situation in terms of our ground station observations and measurements are considered in the next section.

TABLE VII SUMMARY OF TEMPERATURE DATA, JULY 22 - AUGUST 31, 1971

Station	Elev. (ft.)	Setting	Absolute Min.	Absolute Max.	Observed Range	MeanMin. $\bar{x}$	MeanMin. σ	MeanMax. $\bar{x}$	MeanMax. σ	Daily Mean	Mean Max-Min Range
3	7200	W. slope of mt.	24°	56°	32°	33.4	[illegible]	42.[illegible]	7.36	33.1	9.2
10	4000	SW slope of mt.	34	66	32	[illegible]	4.03	51.6	7.04	46.0	11.2
17	4200	arete	34	64	30	42.1	4.62	48.6	6.42	45.3	6.5
26	4900	E. slope of mt.	26	63	42	38.9	4.02	52.2	7.15	45.6	13.3
30	2200	lake shore	32	77	45	42.7	4.71	63.7	6.36	53.2	21.0

Summary of Temperature Data (°F.)
Juneau Icefield, Alaska - British Columbia
July 22 - August 31, 1971

E. COMPARATIVE CLIMATOLOGICAL AND MASS BALANCE INVESTIGATION OF CIRQUE GLACIERS IN THE NORTHERN BOUNDARY RANGE

As noted in the lower inset box in Figure 2 and in the photograph in Figure 5, the Lemon Creek and Ptarmigan Glacier system (Camp 17 area) lies at the southernmost edge of the Juneau Icefield. As such it is in the most maritime of positions. The upper inset box, Figure 2, denotes the Cathedral Glacier position on the most continental edge of the presently glaciated sector of the Northern Boundary Range.

Previous investigations of cirque distribution on the maritime flank of the Boundary Range (Miller, 1960; Swanston, 1967) have been abetted by current studies of inland cirques at this latitude (Jones, 1975; Tallman, 1975). From these it is known that cirques now occupied by ice are close to the glaciation level (Østrem, 1972, 1973). They fit into a series of tandem cirques. On the Alaskan coast, the lowest four are ice-abandoned and the lowest three in the interior are ice-abandoned also. In the Pleistocene relationship to the present glacial position, today's ice-filled cirques lie in the transitional glacial to non-glacial zone. (A fuller discussion of this situation is given in Part X).

The relatively healthy regime of the Lemon Glacier System compared to that of the nine small cirque glaciers on the Cathedral Massif is of special significance to our study. These regime differences are compared in the air photos of Figure 8a and b and relate to differences in geographical position and elevation. The glaciation level lies at 4500 feet on the coast but is at 7000 feet in the interior Atlin region, a fact which is commensurate with today's mean névé-line difference on the south-flowing Taku Glacier (i.e., 2900 feet (880 m) v. Fig. 5) and the north-flowing Llewellyn Glacier (4900 feet (1480 m) v. Fig. 7). Thus it is not unexpected that the mean névé-line between 1971 and 1974 on these lesser cirque glaciers showed a 2000-foot (600 m) difference. The mean névé-line comparisons, including those between the Taku and Llewellyn Glaciers, are noted as follows:

Juneau Sector	Ptarmigan Glacier Névé-line	3700 ft. (1120 m)
	Lemon Glacier Névé-line	3600 ft. (1090 m)
	Taku Glacier Névé-line	2900 ft. (880 m)
Atlin Sector	Cathedral Glacier Névé-line	5800 ft. (1760 m)
	Coliseum Glacier Névé-line	5900 ft. (1790 m)
	Llewellyn Glacier Névé-line	4900 ft. (1480 m)

To delineate more precisely this indicator of maritimity vs. continentality with respect to opposite flanks of the icefield, comparative climatological studies are continuing at each main field research site. Complete details of the regional correlation over the 1971-74 research period are reserved for presentation in later reports.

For our present purpose, specific comparisons are drawn between climatological factors affecting mass balances of the Lemon-Ptarmigan Glacier system near Camp 17 and the Cathedral Glacier at Camp 29. Details of the parameters affecting the Lemon Glacier have been documented in Miller (1972b) and the Cathedral Glacier data (1971-74) have been analyzed by Jones (1975). Pertinent aspects of the Cathedral Glacier study are discussed below, as they relate to

the objectives of this contract. In Part III, the Lemon and Ptarmigan Glaciers are discussed as prototypes of the coastal cirque glacier system.

Climate of the Cathedral Glacier Area

The Cathedral Glacier lies in the semi-arid inland zone. As this is a new research area developed for this study, no meteorological observations had been made prior to 1971. The nearest weather records are from Atlin, 18 miles (29 km) to the north and from Camp 26 on the Llewellyn Glacier, 22 miles to the southeast.

Cathedral Glacier is a small glacier in a closed basin with a single outlet stream (Figs. 8,9). Its unique form has resulted in a complex morainal history that provides quite a complete picture of inland glacio-climatic pulsations during the late Holocene. Most helpful to the present purpose is comparison of the meteorological, glaciological and geomorphic information from this location with that from maritime stations on the southern flank of the Juneau Icefield permitting, in turn, a much more effective analysis of the regional atmospheric behavior.

Significance of Tropospheric Controls *

Because the Juneau Icefield lies at the fluctuating interface between the polar continental anticyclonic high pressure region and the maritime cyclonic region (Fig. 4) during the autumn and early winter season of maximum precipitation, terrain characteristics of the icefield reflect a distinct sensitivity to shifts in the location and changes in intensity of storms. These atmospheric phenomena are fundamentally affected by changes in the general circulation of the atmosphere. The presence and orientation of the coastal Cordillera provide additional control of changing conditions. Elevations sufficient for glaciation amplify the effects and make possible the preservation of a decipherable natural record of long-term atmospheric variations.

The North Pacific/Aleutian low pressure system that dominates the Gulf of Alaska and the southeastern Alaskan coast during the precipitation maximum of the fall is not a static low pressure condition (Fig. 4). Rather, atmospheric conditions in this area are part of the planetary circulation pattern. Long waves of global scale (Wave-lengths of 60° to 180°) steer the overall movements of weather, while shorter waves superimposed on and within the mean long wave positions tend to generate and move individual storm systems. The most probable locations, as well as intensities, of these wave patterns and their storm fronts shift over time. Thus, on a frequency distribution basis, the most probable locations and movements of weather patterns vary over time.

The "Aleutian Low" represents a condition in which the planetary scale high-level westerlies tend to be moving from a wave trough to a ridge as they traverse the Gulf of Alaska. In fall and early winter, low-pressure cyclonic storm systems tend to generate in and move through this area, intensifying as they move from trough to ridge. These storms, moving in a generally northeast direction (Figs. 10,64), bring clouds and heavy precipitation to the coast. As the storms pass the ridge

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* This and the following meteorological analyses (Section F) have been prepared in cooperation with Vernon Jones. A more comprehensive report on the glacio-climatology of the inland sector is presented in his thesis (Jones, 1975).

of the tropospheric wave, they tend to turn southeastward and deteriorate. If they maintain coherence until reaching the base of the trough over the southern 48 states and begin to move ridgeward again, they may reform or even intensify into full-scale rejuvenated storm systems (Fig. 10).

The "Polar High" (Fig. 4) is a condition during which the high-level westerlies tend to move from a wave ridge to a trough. In this, high pressure anticyclones generate in northwestern Canada and move southeastward across the region of Montana and the Dakotas, bringing cold fronts into the central United States and helping to generate new storms as they encounter warm air systems from the south.

Therefore, when we speak of the interface between the Aleutian (sub-Polar) Low and Polar High we are referring to the position of the crest of the planetary wave ridge of high-level westerlies over the Alaskan Coast in the vicinity of the Juneau Icefield. When this ridge is located west of its "normal" position, conditions of continentality move coastward. Thus in the vicinity of the icefield atmospheric pressures become higher, cold air mass characteristics dominate, precipitation decreases, freezing-point elevations lower and ultimately the low-level glaciers expand. When the ridge-crest lies to the east, maritimity prevails farther inland, precipitation increases, temperatures become higher, freezing-point elevations are higher and high-level glaciers expand. There are also changes in direction of storm winds (Fig. 64).

The relationship between relative wave ridge position in the troposphere and precipitation patterns is shown in Figure 10.* In the zone preceding the ridge position (assuming west-to-east air flow), the cell tends to uplift and condense, bringing counterclockwise advection of moist air out of the Gulf of Alaska and a consequent intensification and steering of low pressure storm systems. Air moving from the crest to the following trough increases in anticyclogenesis. With subsidence of the cold and dry upper air, clearing takes place and a clockwise advection of the polar continental air is generated. This situation is illustrative of the condition prevailing at the time of the satellite photo in Figure 4. (Also see Fig. 64a)

From the foregoing it is apparent that a fairly small shift in tropospheric wave location relative to an area on the ground can change weather conditions in that area from a heavy to a light precipitation regime, or vice versa. Thus a persistent displacement of the mean ridge location for an extended period of time will also mean consequent changes in glacier mass balance.

Miller's studies (1956, 1964a, 1972b, 1973) have demonstrated the significance of geographical position and elevation differences in various glaciers in terms of their out-of-phase pulsations in recent centuries. Miller and Anderson (1974) have extended this concept to analysis of longer-range fluctuations in climate-glacier behavior along the coast-to-interior atmospheric gradient discussed earlier in this report. Tallman (1975) and Miller (1975b) have related this further to Pleistocene and Holocene glacial fluctuations in the Atlin and Taku regions. A review of this topic is incorporated at the end of this report to reveal how important an understanding of atmospheric

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* Figure 10 is redrawn and superimposed on an outline map of North America after O'Connor (1963).

controls is with respect to glacial behavior and terrain evolution in the Holocene and Neoglacial that is a basic concern of this study.

The glaciers of the Juneau Icefield have been noted to be uniquely effective recorders of climatic change (Field and Heusser, 1953). They not only provide a means of correlating fluctuations with pin-pointed dates in the historical record, but present paleo-climatic evidences covering periods far earlier than any legends of man. An understanding of the chronology that these glaciers have transposed into physiographic features also provides possibilities for extrapolating future conditions.

Thus the climatological studies across the Juneau Icefield, in which the Lemon and Cathedral Glacier research programs are a fundamental part, have important time-dimensional as well as geographical significance. In view of the present and pending environmental, energy and food crises, all of global dimension, this kind of research with its implications for man to understand and effectively to use climatic trends assuredly is paramount to pursue.

F. DATA ANALYSES OF REGIONAL METEOROLOGICAL RECORDS *

This section deals with the presentation and analysis of some of the typical glacio-meteorological data related to the study area, highlighting records from key locations in the station network on the continental flank of the Icefield and from permanent weather observing sites both inland and on the coast.

For the inland sector, existing daily temperature and precipitation records from meteorological observations at the Cathedral Glacier station (Camp 29) are compared with those at the Atlin station (Camp 30). Although these daily records are available only for the summer field seasons of 1972, 1973 and 1974 (since establishment of the Cathedral station), they provide some basis for comparison between this sector of the research area and Atlin, 18 miles to the north.

The meteorological records for Atlin covering the past half century are then examined. Because the Atlin record is incomplete, a regional data base for this continental interior area is synthesized, using data from Atlin, the Whitehorse Airport and Carcross, Y.T. (Figure 2). Similar long-term records for the Juneau coastal area are then synthesized from Juneau and Juneau Airport records. Following this, the climate on the coastal and interior margins of the Juneau Icefield is compared.

Although no meteorological records exist for the Cathedral area prior to 1972, observations have been made at field sites on the southern flank of the Juneau Icefield for 30 years. Therefore, if relationships between the coast and interior can be established from recent data and long-term regional comparisons facilitated, we have a much sounder basis for analyzing past, present and future glacio-climatic trends in this region.

Also, as previously demonstrated, a sharp gradient of continentality exists between the coastal sector and the immediate interior. In the preceding section it was noted that this gradient is the area over which the tropospheric pressure ridge fluctuates. Therefore similarities and differences in climatic trends across this region have considerable synoptic significance.

Short-term Meteorological Records at a Prototype Site

Sources of Data

During each summer field season detailed meteorological observations were taken every three hours (0600 to 2100) at each occupied station on the icefield. Thus, except for unavoidable problems such as equipment breakage or malfunction, or field operations which prevented the presence of an observer at the scheduled observation time, complete synoptic meteorological records are on file. For Camp 29, the data cover the periods:

1972: July 1-September 19
1973: July 2-September 18
1974: July 5-September 12
1975: June 22-September 15

Similarly, records are available for the Atlin station, obtained by the same standardized procedure and giving a valid comparison of observations between these two stations.

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* In collaboration with Vernon K. Jones (1975)

Analysis

Comparison of mean daily temperatures is shown for both stations for the first three years (1972-74) of the Camp 29 record (Fig. 11). Temperatures generally followed the same trends. Although Camp 29 temperatures ranged from 15°F lower to 4°F higher than Camp 30, during the summer field season Camp 29 was usually about 5° to 7° cooler.

Precipitation at the two stations is shown in Figures 12a and 12b spanning the period of record for Camp 29. Over these three summers, Camp 29 had 12.03 inches of precipitation. Camp 30 had 12.09 inches.

Despite the similarity in the total amount of summer rainfall, there were many discrepancies within single storms and on single days. It is noted that precipitation in this region is almost entirely due to the passage of low-pressure systems rather than to convective instability. In such a situation sharp discrepancies are observed even between adjacent stations in a dense network of rain gages.

From this comparison of daily precipitation, it is concluded that rainfall records at the Atlin station provide a fair approximation of summer rainfall at the Cathedral research station. However, it is not safe to extrapolate this conclusion to the remainder of the year, due to vastly different atmospheric conditions during the colder seasons. As an example of variations in winter conditions refer to the April 1 snow depth data for Atlin and Log Cabin noted later in this section.

Long-Term Meteorological Records

Sources of Records and Data Preparation

Useful in this study are the official government weather records obtained at first-order stations at the Juneau Airport and Whitehorse beginning in 1942 and 1943 respectively. Records also have been maintained at the stations noted below by non-professional cooperative observers over varying periods of time. In this listing, parentheses indicate a fragmental or incomplete record (see Figures 1,2 and 3 for site locations).

Atlin, B.C.	1906-45; 1967-71; (1972-75)*
Carcross, Y.T.	(1909-13); 1914-26; (1935-40)
Engineer Mine, B.C.	1926-28; (1929); (1975)
Whitehorse Airport, Y.T.	1943-75
Juneau (City) Alaska	(1905-11); 1912-69; (1970-71)
Juneau Airport	(1942-45); 1946-75

Complete records do not exist for any single station in the region. Therefore, in order to have data covering an adequate time span it has been necessary to

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* It is noted that the JIRP Atlin station has maintained full synoptic records between 1969 and 1975. These are in more detail and more consistent than the government agent's records referenced here.

develop composite and representative data sets from portions of records from different stations. In doing this, temperature and precipitation data have been adjusted to common equivalents to avoid introducing apparent variations that could be the result of combining records from different locations in the region. For example, on the Alaskan coast if precipitation data are obtained from Juneau City records covering the period up to the time when the U.S. Weather Bureau Office was established at the airport, and data are also obtained from the Juneau Airport records, it could appear that the Juneau area annual precipitation dropped abruptly by 38.5%, which was not the case.

By converting properly sequential records, however, for each region to long-term equivalents, it is possible to develop a continuous data base for both the coastal and continental interior regions covering the years 1907 to present. This particularly pertains to the comparison of mean annual temperatures and total annual precipitation.

To obtain a set of mean annual temperatures that are representative of the immediate interior region near the Cathedral Glacier, the means of stations with overlapping records were compared. It was found, for example, that A lin and Whitehorse Airport temperatures are very comparable, while Carcross temperatures average about 1°F lower.

It was also found in determining representative coastal temperatures that for the period of common record the Juneau Airport temperatures averaged 2°F lower than the Juneau City temperatures. Therefore, where both figures are available, the mean was used; otherwise the Juneau City temperature was used minus 1°F or the Airport temperature plus 1°F. The consistent discrepancy in mean annual temperatures may be due to: (1) rooftop exposure of the city instrument shelter; (2) the city "heat island" effect; (3) the location of the airport downstream from the Mendenhall Glacier valley, exposing it to katabatic wind effects; and (4) the airport's position, exposed to the open water of Lynn Canal.

It is also noted that since 1959 the average mean annual temperatures of all stations in southeast Alaska followed the Juneau area temperature trend, but with values consistently about 1.5°F higher.

Temperature Analysis

Temperature records for the immediate continental interior, representing available data nearest to the Camp 29 research locale and for the Juneau area near Camp 17 were analyzed. The data were plotted graphically and processed by computer, and the statistics developed from computer analyses were tested for level of significance.

Frequency distributions were plotted for the mean annual temperatures for both regions (Jones,'75)Several observations are noted from this record and from the accompanying computed averages.

(1) There is no overlap between the interior range in mean annual temperature of 25 to 36°F and the coastal range of 38 to 45°F.
(2) The interior averages vary more, with a total range of 12° F, compared with a range of only 8°F on the coast.
(3) Mean annual coastal temperatures averaged 10.5°F. warmer.
(4) The coastal temperature range is continuous and terminates sharply while the interior has outlying extremes.

(5) The range of mean annual coastal temperatures is skewed to the right, while that for the interior has relatively few years below the modal number.

Actual mean annual temperatures for the continental interior region and for the coastal Juneau region are shown in Figure 14. Two characteristics stand out:

(1) Sharp oscillations in mean annual temperature from one year to the next. This is more pronounced in the interior.
(2) Year-to-year temperature oscillations on opposite sides of the Juneau Icefield are in close agreement. In other words, the temperatures on an annual basis in both regions, though quantitatively different are quite parallel and in-phase.

Since year-to-year changes are so extreme, it is necessary to smooth the data in order to determine if trends actually exist. Three methods were used: a weighted moving average; a linear regression; and a polynomial regression.

Weighted Five-year Moving Mean

In this method a weighted average was applied, of the form:

$$\bar{x} = 0.1\ (x_{n-2}) + 0.2\ (x_{n-1}) + 0.4(x) + 0.2(x_{n+1}) + 0.1\ (x_{n-2})$$

$\bar{x}$ represents the mean, x the mean annual temperature of a given year, and n the central year of the particular five years being averaged.

The purpose of the weighting was to avoid the Slutsky-Yule effect (Mosteller et al, 1973). When an un-weighted moving average is applied to a series of random numbers, an apparent periodicity is generated. In research involving a series of temperature data, such an effect can generate a spurious apparent cyclicity or can mask a real pattern which does in fact exist. The five-year weighted moving average is shown in Figure 14, with inland and coastal areas plotted.

Inspection of the data shows a cyclical upward trend in S.E. Alaskan coastal temperatures until the early 1940's, followed by a downward oscillation since that time. Periods of oscillation are roughly between 13 and 17 years in length. Again, the periodicities in the interior and coastal sectors agree.

Magnitudes of the oscillations in the weighted five-year means are also affected by two factors: (1) the level of extreme temperatures and (2) the persistence of departures from the central value. While 1927 was the peak year for temperature it was between two below-normal years. Persistence of above-normal annual temperatures from 1940 through 1947 gave a very positive level in the five-year weighted average. Similarly, temperatures consistently below normal for 1948 through 1951 without a warm year between gave a low point in the moving average.

Linear Regression Analysis

A relatively simple method (Blaylock, 1960) of determining the rate of change in a variable over time is a simple linear regression of the form:

$$y = a + bx$$

Application of the appropriate statistical formulae provides values for a and b in which a is the point at which the regression line crosses the y axis,

b represents the slope of the regression line and x is the value of the independent variable (in this case the year) for which the dependent variable (the predicted temperature) is to be determined. For our purposes, x is the first year in the series and x_n is the last year. The slope of the line (b) and the value of y when x = 0 provide information as to the magnitude of the change. This is applied to net change in the value of the dependent variable from the beginning to the end of the period. A dependent variable could rise to a high value in the middle of its range and then return fairly symmetrically to its original value, and a linear regression could indicate that no change had occurred.

Regional temperatures peaked in the early 1940's after oscillating upward since the beginning of the climatic record. After that time they have oscillated downwards. Therefore the linear regression analysis as applied separately to both coastal and interior records is broken into two parts. The first subset for each location includes the years from 1907 to 1943 and the second subset covers the period from 1944 to 1973.

A caveat is in order when applying linear regression techniques to oscillatory data such as these temperature records. The slope of the regression line may be conditioned by the portions of local oscillations on which the regression line ends. In other words, starting at the low point of a cycle and ending at the maximum of a later cycle can exaggerate the slope, while starting at a high point and ending at the minimum of a later cycle will minimize the slope. Where the range of oscillation is greater than the long-term change in trend, as in the case with our data, the latter error may reverse the sign of the slope entirely. Results of the linear regression analysis are shown in Figure 13. Coefficients were computed by a scattergram computer program.

Polynomial Regression Analysis

Since plotting of the five-year weighted mean temperature data showed an apparent quasi-periodicity, a polynomial regression computer program was applied to the temperature data from the coastal region and separately to the data from the interior, for the period from 1907 through 1973. The predictive equation for the polynomial regression is:

$$y = a + b_1x + b_2x^2 + \ldots . b_nx^n$$

In this equation a is again the y intercept of the equation and b is the mean slope of that segment of the line. The unit x is the value of the independent variable, while y is the predicted value of the dependent variable (temperature) for that value of x (the year). The exponent of the value x indicates the degree of the equation in that segment (Blaylock, 1972). The degree of the equation may be defined as one less than the number of times the value of the slope changes signs. For example, if the slope of the segment b_1x is positive, the slope of the segment b_2x^2 will be negative and the slope of the segment b_3x^3 (third degree) will be positive.

A polynomial regression provides the equation and hence the plot of a line whose formula contains the fewest degrees of x (in essence, changes in sign of the slope of the curve) which are necessary so that the improvement of an additional degree does not give a significant improvement (decrease) in the sum of the squares of the deviations of the observations from the plotted polynomial regression line.

Results of the polynomial regressions, specifically concerning temperature fluctuations and trends, are shown in Figure 15. This analysis, too, showed a <u>regional temperature maximum in the early 1940's and a very sharp drop in recent years.</u>

The value plotted for a given year is affected by the observed values of those preceding and following the given year. As the computer print-out plots show a tendency for the unsupported end value to go to extremes, the valuos computed for the first and last three years were deleted.

Since we are dealing with two different series of oscillatory trends of opposite slopes, the polynomial regression was applied separately to the two segments of data. The results agree remarkably well with the five-year weighted moving averages for both the coast and the interior temperatures (Fig.14). Again, unsupported extreme end values were deleted, although this caused a gap in that part of the polynomial regression graph representing the early 1940's.

Regional Temperatures

Observations and Conclusions

The following observations and conclusions regarding regional temperature patterns emerge from the foregoing analyses:

(1) Annual mean temperatures in the interior are consistently colder than on the coast. Interior means range from 25°F to 36°F, with a mean of 30.75 °F from 1907 through 1973. Mean annual temperatures in the coastal Juneau area range from 38° to 45°F, with a mean of 41.4°F.

(2) Interior temperatures closely parallel those of the coastal region.

(3) Visual inspection of annual mean temperatures in the 67 years of record reveals a pattern of about four quasi-cycles with a mean length of about 17 years.

(4) A five-year centrally weighted moving average reveals and corroborates the above patterns more clearly. The first three oscillations are consecutively more pronounced, with the third peaking in approximately 1944. The fourth is of notably less magnitude.

(5) The pattern of the five-year means is more erratic after the 1950 minimum.

(6) Temperatures have been in a downward skid since 1959, the cooling trend hav[illegible]opped even more sharply since a minor respite in 1967-69.

(7) Patterns are somewhat erratic for accurate prediction, but a subjective projection of the trend suggests that mean annual temperatures twenty years from now will be far below any yet experienced in this century.

Regional Precipitation Analysis

Comparative annual precipitation totals for the coast and interior region are shown in Figure 17. It was necessary to use an expanded scale for the interior data to allow effective visual comparison. Precipitation in this interior region averaged 11.03 inches per year over the period 1907 through 1974. This is only 12.8 % of the 87.1 inch yearly average at Juneau for the same period, clearly illustrating the sharp gradient of continentality across the Juneau Icefield.

The validity of using Juneau City precipitation values as representative of the coastal sector is supported by the close agreement in trends and in means between Juneau City (93.1 inches) and all southeastern Alaskan stations (92.0 inches) for the period 1959-73. In a statistical analysis of precipitation data covering 64 stations along the Alaskan coast for the period 1907-1962 Miller (1963) demonstrated unusually high correlation coefficients of 0.80 to 0.94, strongly supporting this conclusion.

In order to smooth the data and more clearly reveal trends, five-year weighted moving means were applied to annual precipitation records using the same formula as with the annual mean temperatures. Analysis of Figure 18 reveals no clear-cut trends over time, with the exception of a relatively dry period along the coast around 1910. A roughly approximate seven-year periodicity appears along the coast, except for the 1910-17 period. If the 1965-72 period does appear, it has a sharply restricted maximum. The existence of a seven-year pattern is much less clear in the interior. Oscillations in precipitation in the interior show some slight tendency to be in-phase with patterns along the coast during parts of the record, while other parts show no trend toward parallel variation. There appears to be no tendency toward diametrically out-of-phase variation. Parallel phasing is more pronounced in pattern although not in magnitude toward the end of the study period.

Regional Temperature and Precipitation Relationships

The work of Anderson and Miller (Anderson, 1970 and Miller and Anderson, 1974) has shown some relationship between temperature and precipitation over the time of the Holocene, a time scale on the order of 10^3 years. In the Taku District near the coast, warmer temperatures have been associated with drier periods. In the interior Atlin District, warmer temperatures were associated with wetter conditions and increased storminess. The latter condition would suggest an eastward shift of the tropospheric ridge, with inland penetration of dominating low-pressure cyclonic conditions. From a study of the Wisconsinan moraine sequence in an interlobate sector of the Atlin region over a longer time span, Tallman (1975) has presented some evidence that this shift has taken place during the larger climatic changes of the Pleistocene.

Five-year weighted moving means of temperature and precipitation were plotted for the coastal region and for the interior(Figures 19,20 shown together for comparison). As was the case with temperatures, existing patterns broke down after the peaking of temperature trends in the early 1940's. Prior to the 1940's, temperature and precipitation had a strong tendency to be diametrically out-of-phase. Since that time, there has been a tendency for the patterns to be in-phase along the coast.

Figure 21 shows the pattern of snowfall accumulation at the Juneau Airport for 1943 through 1974, revealing a seven-year periodicity and a general upward trend. Total precipitation displayed a seven-year periodicity from 1917 to about 1941, when the pattern broke down. A short six-year pattern reappeared briefly from 1947 to 1957, after which the pattern again broke down. When the five-year snowfall mean is compared with five-year temperature means little relationship appears. Similarly, five-year average snowfall bears little relationship to total precipitation trends which lag one to two years behind the temperature trend.

In addition to the change in patterns of temperature oscillation and the phase reversal of the temperature-precipitation relationship after the early 1940's, the most recent behavior of temperatures emphasizes the anomaly. An upswing in the 13 to 17 year temperature oscillation, which past patterns suggest as being at a maximum about 1975, has either not clearly begun or else appeared only faintly about 1969 followed by a sharp drop in temperature. Annual temperature means over the next several years will be critical in suggesting what paths, anomalous or otherwise, temperatures may take and the magnitude of the change.

In terms of human catastrophe, extreme climatic changes have been painfully obvious on a global scale in the past seven to ten years. Many causes have been suggested to explain what in the popular press has been called the "ominous" changes in world climate. The list of possible causes includes particulate and chemical air pollution, increase in carbon dioxide, effects of atomic bombs, over-grazing, agricultural land and forest-clearing fires, and most recently aerosol cans...with the chlorine in their Freon propellant apparently reacting with and destroying the protective shield of ozone in the high atmosphere. It may be that man has already set in motion the forces for his own doom (Fortune Mag.,Alexander,'74). It may be, on the other hand, that his pollutants may have no real effect in comparison with natural processes on a global atmospheric scale (Willett, 1975).

It may be that dominance has shifted from one atmospheric control to another. Or that the atmosphere is merely returning to more normal conditions after a 40-year period of abnormally favorable weather, so that we need to look for the cause of the initiation of the "anomaly" rather than for a cause for its effects.

Solar Cycles, Temperature Trends and Glacier Regimes

Many investigators have noted relationships between weather and the occurrence and periodicity of solar disturbances, i.e., electro-magnetic (wave) radiation and charged particle (corpuscular) radiation revealed by sunspot variations (Miller, 1956, 1972; Willett, 1975 and other publications). To explore this relationship the five-year weighted moving mean of annual average temperatures for the coastal region is plotted against the Zurich average annual sunspot number (R_z) as shown in Figure 22.

Lack of a discernable relationship between temperature and sunspot data is supported by a statistical correlation of 0.024. Application of temperature lag times to account for the oceanic heat sink effect provides an even poorer fit. Similarly comparison of short-term sunspot numbers with precipitation patterns, both on the coast and in the interior, show no closer relationship. It is concluded that any existing short-term relationships between temperature and the 10.8 year solar cycle are too complex to discern readily and hence if such do exist the proof will require more sophisticated analysis.

Because of these difficulties, another approach was taken. For example, mid-18th century lichenometric dates on major Little Ice Age glacial advances in the study area were used as benchmarks for inferring postulated glacial regimes. This pertains not only to the Cathedral, Lemon and Ptarmigan Glaciers but also to the known recent behavior of other glaciers on the Juneau Icefield (see Parts X and XI). Known and inferred temperature patterns in the Juneau Icefield region were plotted. Finally, for comparison, the Zurich sunspot numbers from 1749 through 1974 (Appendix C). were plotted on the same time scale. The results are shown in Figure 22.The following relationships are apparent:

(1) Periodicity of glacier behavior is very close to 90 years.

(2) Glacier terminal behavior lags behind temperature patterns by about a half cycle (45 years). The glacier advances during a period of rising temperatures and retreats during the cooling portion of the apparent 90-year cycle.

(3) Short-term temperature changes do not appear to be related to short-term (10.8 year) solar cycles. The inertia and lack of resolution of finer climatic perturbations in the system may be too great to allow detection of changes at this scale.

(4) During the Little Ice Age, glacier advances appear to have occurred during and after the minimum portion of the 80-90 year sunspot cycle. Retreats tended to occur during and after the maxima of these "long" cycles.

(5) Temperature peaks tend to <u>lead</u> the 80-90 year solar peaks by about 15 years. This bring up a tantalizing question concerning what has been considered by Miller (1972b) to be a 15-year cooling lag attributed to the heat sink effect of the North Pacific Ocean.

If indeed there is a close solar-temperature relationship, over the time scale of this study there may be a slight drift in the relationships of temperature and glacier regime in comparison to patterns of solar behavior. In the absence of known causal factors, it is suggested that we may be witnessing the in-phase coincidence of two unrelated cycles of slightly differing lengths. Thus not on a scale of the three "long" cycles plotted in Figure 22 but of ten, twenty, or fifty, the relative patterns could be completely reversed.

Relationship to Global Atmospheric Circulation

A hypothetical causal mechanism was suggested 20 years ago by Miller (1956) which is counter to implications of the idea in the preceding paragraph. Solar activity affects the emission of energy from the sun partly in the form of particulate matter (corpuscular radiation). These charged particles are to some extent channelled to the magnetic field, as has been considered in previous JIRP publications (also see Miller, 1963, 1972). The resulting energy-input causes fluctuations of the wave-length of the circumpolar westerlies, changing the location of the tropospheric ridge along the coastal mountains of Alaska-Canada. The relative locations of the mean low pressure systems in the Gulf of Alaska (Aleutian Low) and areas of anticyclogenesis in the interior (Canadian Polar High) would thus be a reflection of ridge location. In turn , temperature and accumulation changes in the Boundary Range would be affected. Such changes

in temperature and precipitation regimes are integrated in the behavior of glaciers and ultimately in the landforms they produce. But some inconsistencies might appear to remain in this mechanism. For example, we should expect increased energy to increase the activity of the westerlies. But would not the increased turbulence and meridionality, being due to increased thermal contrast, mean colder polar areas, with <u>less</u> energy (Jones, 1975)? The answer lies in apprecia-tion of the fact that we <u>cannot</u> restrict our view to one corner of the system. We must continue to consider the whole...i.e.. the g[illegible]eric circulation.

<u>The Phasing of Temperature and Precipitation Periodicities</u>

Short-term comparison of daily meteorological observations at the Cathedral Glacier (C-29) and Atlin (C-30) research stations during recent summer seasons reveals that at Camp 29 the average temperatures are 5 to 7°F cooler, but that total precipitation is about the same despite day-to-day differences. Records for the winter seasons are not available at Camp 29, so comparison is restricted to the summer months. A summary of mean annual temperature and precipitation between the coastal area and the interior is given below.

Summary of Temperature/Precipitation Data

	Coast	Interior
Range of Mean Annual Temperatures (°F)	38-45	25-38
Amount of Range in Mean Annual Temp.(°F)	8	12
1907-73 Average of Mean Annual Temp. (°F)	41.4	30.75
1907-73 Average of Annual Precipitation	87.1 in.	11.03 in.

The wider range in mean temperatures in the interior reflects the continentality gradient. While mean annual temperatures fluctuate sharply from year to year, the fluctuations are usually parallel between the coast and the interior. Rain-fall patterns between the two regions are much less in phase. The extent of phasing that occurs varies considerably over time.

The analytic techniques applied in this study, including graphic analysis and linear and polynomial regressions, reveal an approximate <u>17 year periodicity in temperatures</u> and a less distinct <u>7-year pattern in precipitation</u>. On a longer term basis, temperatures have generally moved upward from 1907 (the start of the record) until the early 1940's. Since then they have moved downward, with the rate of decrease accelerating over the past decade.

The most significant observation developed in this analysis is that the nature of atmospheric relationships has changed with or shortly following the peaking of mean annual temperature trends in the early 1940's. As patterns of temperature oscillations changed, the phase relationships of temperature and precipitation have nearly reversed. Also, temperatures in the past five to ten years have dropped much more sharply than earlier patterns would suggest, and the tentative outlook is for this trend to continue despite short-term upswings. The question arises as to whether the sharp temperature drop since the early 1960's is a natural climatic trend or an anomaly. Field data on the Juneau Icefield Research Program over the next five years should provide a clearer picture of the short-term situation and a better idea of expected trends and basic causal factors.

PART III INVESTIGATIONS IN GLACIO-HYDROLOGY

In line with our objectives to study processes in their relation to terrain evolution, several special studies are discussed in the following pages. The first of these describes and discusses the processes of a self-spilling lake on the Lemon Glacier and its affect on Jokulhlaups (glacier bursts) in the Lemon Creek Valley (Miller and Asher, 1975). The second treats the general hydrological characteristics of the Lemon-Ptarmigan glacier system (Brush, 1972) and a third considers a preliminary assessment of the Cathedral Glacier hydrology, where such fascinating problems as the evolution of sine-generated curves in supra-glacial streams are found (Nishio and Miller, 1972). A fourth reports on the regime of ephemeral supraglacial lakes, emphasizing the detection of thawing firn on the Juneau Icefield using ERTS satellite imagery (Bugh, 1975). (This technique has large potential for monitoring future changes.) A fifth discusses the thermal erosion of fluted surface channels on the Vaughan Lewis and Gilkey Glacier systems (Pinchak, 1972, 1975).

Other shorter studies are reported referencing the field work of Dr. Edward Little (1972) and Dr. Gorow Wakahama (1972). Special assistance in these studies is acknowledged to Richard Heffernan, John McCracken, Clarke Petrie, Gregg Lamorey, Paul Willis, Howard Langeveld and other field assistants and participants on the Juneau Icefield Research Program between 1971 and 1974.

The individual reports within this discipline of special environmental concern follow.

A. A SELF-DUMPING GLACIER LAKE AND RESERVOIR CAVE ON THE LEMON GLACIER*

Ephemeral self-dumping water bodies impounded in and on glaciers (v.Fig.78) have been the subject of much interest in recent years (Stone, 1963). Such a supraglacial lake occurs at the head of Lemon Creek Glacier**, at the extreme southwestern edge of the Juneau Icefield and but a scant 5 air miles (8 km) northeast of Juneau (Fig.23). What distinguishes this lake is a uniquely associated glacier cave system. The accessibility of the lake and its englacial cave (Fig.24) permit continuing and detailed measurements to be made of key hydrological and climatic factors affecting this glacier's regime.

Although the literature on glacier caves has been increasing, a consensus terminology has not yet been developed (Halliday, 1966; Peterson and McKenzie, 1968; McKenzie, 1969, 1971; Kiver and Mumma, 1971; Miller, 1972c). Because of the unusual characteristics of the cave system, in this presentation it is referred to as a composite cave. Its extensive englacial tunnel system was surveyed in August and September 1972 and again in August, 1973. The resulting maps are shown in Figures 24 and 25.

Hydro-environmental Implications

Liquid water associated with glaciers and with their supraglacial and englacial reservoirs occurs in large volumes, and in many cases can represent a useable hydrological resource (U.S. Geological Survey, 1971). In Europe some glacierfed streams have been tapped for hydro-electric purposes (Østrem, 1972b). Especially in Norway, France, and Switzerland, the need for water resources planning has dictated that glacio-hydrological data be quantified. With such an aim, long-range measurements have also been conducted in many other countries with respect to the mass balance of glaciers during the International Hydrological Decade (1965-74).

In some glaciers, supraglacial lakes are uniquely sensitive measures of shortterm regime changes (Haag, 1969). Rapid and catastrophic drainage, as during a Jokulhlaup*** (Thorarinsson, 1939), quickly covers flood-prone areas downvalley with water moving at high velocity (Rothlisberger, 1971, 72) When this occurs, any structure or habitation in the pro-glacial area can be in danger. Summer time floods of this kind are found on the Juneau Icefield. Prototypical examples can be observed in the peripheral sectors of the Berner's Bay Trench and of the Tulsequah and Taku River valleys (Fig. 2), where large-magnitude catastrophic glacier-bursts occur. Every year these floods have been of much concern to mining, road building and engineering operations in this region (Fig. 79) (Kerr, 1934,36; Miller, 1952,63; Marcus, 1960; Post and Mayo, 1971). As more development takes place in Alaska, it will become increasingly vital to understand and predict this phenomenon and to survey changes in related supraglacial and englacial drainage and reservoir systems. Between 1965 and 1974, monitoring of the Lemon Glacier drainage has been carried out as part of a continuing longterm glacio-hydrological study of this representative glacier system (Miller, 1972b).

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* This section, prepared by Maynard M. Miller and R.A. Asher, is included with the permission of the Journal of Glaciology. (Also see preliminary report by Asher et al (1974) and Appendix A, p. I, and photos p. 91 a, b)

** Locally known as the Lemon Glacier

*** Icelandic for "glacier bursts", a spontaneous or catastrophic hydrological surge.

Morphology and Drainage of the Englacial Cave

The supraglacial lake on Lemon Glacier is situated in a marginal moat at the head of the glacier (Fig.23). This impounded water body is known as Lake Linda (Asher, 1970; Smithsonian Institute, 1971). The western cave entrance was first discovered in 1969. In the summer of 1972, after Lake Linda had completely drained, this cave entrance was again exposed at the deepest (southwest) end of the lake and another one opened up at its eastern end at Lynn Falls (Fig. 24). These entrances permitted access for the complete mapping of the cave section lying between these openings. The exploration revealed an elongated tunnel with several distributary extensions beneath the ice, the total system of which throughout most of the year serves as an extensive reservoir for water impounded below the level of penetration of the annual cold wave from the glacier's surface.

The main body of the cave is a half kilometer long, connecting Lake Linda to Lynn Falls. The axial distance is 1470 feet (451 m), with the straight-line distance between entrances being 269 feet (82 m) shorter. In the last two years, the cave has been accessible in late summer, after the entrances opened and the lake had fully drained. During the rest of the year, the cave has remained either partly or totally flooded with its entrances buried by winter snow. Partial flooding is suggested because of the abundant stalactites, ice stalagmites, ice draperies, sublimation crystals, hair ice and other glacial speleothems found in the cave system soon after its entrances opened.

The cave reservoir is, in effect, a sub-glacial extension of Lake Linda. In recent years there has been no evidence that water from Lake Linda has drained via this tunnel through the moraine-capped headwall col into Salmon Creek Valley to the south (Fig. 23). In every summer over the past decade, Lynn Falls has become dry before onset of the lake's drainage. Usually the drainage requires several days to run its course, but on occasions the self-dumping process has been observed to be completed in a few hours. Because the falls, prior to drainage, become dry, the water must pass down-glacier through the several distributary channels in and beneath the ice until it reaches the Lemon Glacier terminus. The straight-line, down-valley distance for this flow is approximately 4 miles (6.5 kms).

During the spring and summer melt-season when there is copious melt-water in the firn-pack, Lynn Falls serves as a release mechanism to maintain the hydrologic level of the headwall area at a relatively stable position. The situation may be likened to that pertaining with an overflow gutter around a swimming pool. The falls drain off water above a certain level, but when the water sinks below that level there is no effect and the waterfall channel becomes abandoned. In the case of sudden release of water from Lake Linda (again as in the swimming pool analogy when a plug is "pulled"),all water goes down the drain. In this field case, the drain is simply the lowest outlet of the glacier's reservoir system.

Surface and Sub-surface Mapping

Although mapping of the Lemon Glacier has facilitated the study of ice movement at the glacier's surface, measurements of englacial flow and basal slip have been difficult to obtain. Such measurements can, however, be made in this kind of tunnel system in sectors where it reaches bedrock. During the 1972, 1973 and 1974 seasons, geophysical investigations of ice thickness and related surveys

of position stakes at the névé surface were made in the glacier's uppermost accumulation zone. Seismic, gravity and magnetic stations were positioned directly over the glacier cave (Hinze, Kellogg, Shaw and Stallwood, 1973), with each located on the surface traverse from survey stations 37 to 43 (Fig. 24). Glacier depths from the névé surface to the tunnel ceiling were as much as 79 feet (124 m).

An IR Distance Meter (accuracy $\pm$ 1cm(.4 in) in 2 km (1.2 mi) and a T-2 theodolite were employed for survey control at the cave entrances and for related névé transects above it. In the 1972 mapping project, tape and compass surveys were made on the tunnel floor, with bench marks established in bedrock. The longest chained section was 182 feet (55.5 m), between stations 17 and 18. The shortest was 8.17 feet (2.5 m) between stations 3 and 4. The survey started at Lynn Falls and ended at Lake Linda. Closure was accomplished by reference to bench marks on bedrock on the western shore of the lake. Surveys were checked by reverse traverse and accuracies obtained within 2 inches (5 cm) at each of the 27 control stations. In 1973 the cave system was remapped, this time using a T-2 theodolite for a complete transect of the tunnel (v. later discussion).

Within the cave,cairns were erected at all numbered survey flags and cross sections sketched at key points. Slope distances between survey stations were measured by tape. Measurements for the cross-sectional plots were obtained using a telescoping stadia rod extended horizontally between cave walls and vertically between floor and ceiling. Map distances were calculated from horizontal and vertical angles taken in both directions at each survey site. Many problems were encountered in using the surveying instruments with flashlights in a black and constantly dripping cave. Bedrock was exposed intermittently on the cave floor, with quantities of rock and ice rubble present elsewhere. Several distributary drains extending beneath the main body of the glacier to the north were explored for short distances but these thinned to impassable dimensions. Ice pillars extending downward from moulins were observed, as well as tectonic (flow) foliation on the glacier's bed. Thrust structures were also exposed on the floors, walls and ceilings. Specific locations of these features are shown in Figure 24.

Two maps produced in 1972 show a longitudinal cross-section and a sequence of lateral plots of the main cave on a planimetric base. These are presented in Asher et al (1974) with the longitudinal plan and cross sections reproduced in Figure 24. On the latter plot survey station 24, at the Lake Linda entrance, is seen to be 16.4 feet (5 m) lower than station 1 at Lynn Falls (referenced in later discussion). On the more detailed planimetric maps (Fig. 24, upper), information on the physical character of the cave is given. A third map produced during the 1973 field season (Fig. 25) , also with longitudinal plan and cross-sections, shows the configuration of a main distributary arm of the cave system at a point originating just below the Lake Linda entrance of the main tunnel.

To facilitate re-mapping the cave, permanent survey markers were implanted on the floor in 1972. At several stations, crosses were chiselled in rock and at four locations meter-long steel pipes, appropriately coded, were driven into dense glacier ice.

The survey control could be enhanced by direct mapping of a common point on the surface of the glacier and directly beneath, inside the cave. This should be done via a bore-hole, although possibly a moulin could be used. Since field conditions have precluded drilling a hole and the moulins have not been sufficiently exposed in the last two years, it has been essential to extend the survey control into the cave from the entrances. The northeasterly trending extension (Fig. 25), which was covered in 1970, was found to be open enough in 1973 to explore and map. In 1974 it was not until early August that the reservoir tunnel first opened enough via drainage of Lake Linda to allow re-mapping. During this project, a total of 17 of the 27 flagged stations from 1972 were found in place. Ten flags could not be recovered, five having been lost in the water-swept entrance section of the cave. All the steel pipes imbedded as englacial bench marks in 1972 were found in 1973, having remained intact in the dense glacier ice on the tunnel floor. Glaciological aspects of the 1973 study compared with those of 1972 are considered later.

<u>Ice Speleothems and Sublimation Features</u>

The Lemon Glacier is geophysically Temperate. On its highest névé at 4200 feet (1,243 m) the cold wave by early March has penetrated between 10 and 15 meters beneath the glacier surface. Every spring this cold wave dissipates and returns the glacier to 0°C, usually by the middle of May (Miller, 1972 b). The character of thermal penetration is revealed by data from thermistors drilled into the ice from the glacier's surface. Because of the temporary sub-freezing conditions of winter and spring, much cave coral develops from dripping and splashing water and many sublimation crystals form over the cave's floor, walls and ceiling. Surprisingly some of them survive well into summer. By August, however, within the cave the temperatures of ice, air and water equalize at close to 0°C. It is probable that in some summers slightly sub-freezing conditions pertain within some sectors of the ice, as has been described in a Temperate to sub-Temperate glacier cave at a comparable elevation in Norway (McCall, 1953).

As our mapping progressed, many of the frost features melted out, especially in the entrance areas between survey flags 1 through 4 and 21 through 24, where the exposure to warmer outside air over a few days produced notable changes. Near survey stations 16 and 17, some crystal boundaries within the glacier were observed as beautifully etched at the surface in dense bubble-free glacier ice, with individual crystals measured up to 10 inches (25 cm) in length. These also showed stress orientations parallel to foliation in the ice at places where the overlying glacier was upwards of [illegible] meters thick. Also ornate sublimation crystals were seen and much hair ice was found to extend up from the cave floor in delicate "spaghetti"-like shapes (helictites). Over large areas these crystal forms stood on end like bundles of straw to heights up to two inches (3 to 5 cm). Others were planar in form, some represented by unusually well-developed Tyndall figures, most of which melted away in the first few weeks following the freezing season. Preferred locations of ice sublimates on the floor led to our belief that, indeed, sub-freezing temperatures are retained beneath some portions of the cave. Organic material, some heath and grass and black organic sludge, was also found incorporated in dense water ice between stations 17 and 18. These remains may be pre Little Ice Age in origin, which , if true, has significance with respect to the initiation of the cave.

One of the most striking features were the ice pillars(columns) near the bottom of moulins and extending down through the cave ceiling. These varied in size from 8 feet by 1 foot in diameter (2.5 m by .35 m) to 33 feet by 4 feet (10 m by 1.3 m) (Fig. 11). The columns were often deformed and twisted and showed the

effects of stress. They were the result of freezing water percolating down from the restricted base of moulins or other percolation channels through which drainage into the tunnel had been achieved (note ice column locations on map in Fig. 24). Some pillars were found to be hollow. If after formation they remained in contact with the ceiling and floor, they became deformed by shear stress from glacier flow.

A Sub-Glacial Reservoir

The considerable size of the tunnel system adds about 40 per cent to the volume of water impounded in Lake Linda at its maximum filling. This substantially increases the liquid water capacity of the headwall sector of the Lemon Glacier. The slow drainage of Lake Linda during the summers of 1969-74, even when the lake surface rested below the outflow level of Lynn Falls, corroborates the idea that this system considerably extends the storage capacity of the lake. Also it should be noted that some impounded water, as well as the uninterrupted dripping off the ceiling, comes from side-wall drainage and surface melting of the overlying firn. The vertical extension of moulins and crevasses is noted in the cross-section map where ice columns have developed. The horizontal and vertical locations are indicated in Figure 24. This is also well documented in Miller and Asher (1975).

Though the 1969-74 period of this study has revealed a normally slow outflow from Lake Linda, it is of interest that in 1967 sudden drops in the lake level were observed and recorded (Thorndike and Twomey in Miller, 1972). (Another such rapid drop was observed from August 2 to 5, 1975). The former is reflected in the hydrometric plot of Figure 25 which shows the daily discharge of Lemon Creek at our stream gage site, 1.5 miles (2.5 km) down-valley from the glacier snout (Fig. 23). In this hydrometric plot, covering the full 1967 ablation season, the occurrence of three Jokulhlaups is suggested...i.e., on July 29, August 20 and September 16. At these times Lake Linda was observed to drain rapidly, with the drainage in each case completed in less than 12 hours. Also at these times, no increase in water flow was observed at Lynn Falls so the drainage is concluded to have been sub-glacial and responsible for the above-noted anomalous peaks in the Lemon Creek hydrometric curve. The study of rainfall histograms at the bottom of Figure 26 suggests that these peaks cannot be correlated with significant abnormalities in precipitation.

The cause of hydrological surges is not clear, but it presumably relates to unplugging of sub-glacial channels down-valley from the headwall tunnel. It may also relate to sudden adjustments to stresses within the ice, resulting in opening of new passageways. The question of unplugging in ice-dammed lakes has been theorized by Glen (1954) to relate to development of a critically limiting hydrological head. Comparison of continuous hydrometric and meteorological records with sequential changes in the lake-water level (volume variations) and pressure fluctuations in its tunnel extension could clarify the causal factor (Rothlisberger, 1972) or at least throw more light on the frequency of Jokulhlaups. It is reiterated that in the four summer seasons of 1971-74, the drainage has not been observed to be spontaneous or catastrophic. It is further noteworthy that these four summers have been unusually cool, with névé-lines tending to be lower than normal.

In mapping the glacial stratigraphy and ice/bedrock contacts, old dense glacier ice was found on the north side of the cave. Here it is consistently overlain by cleaner and younger bubbly glacier ice. In a few areas, older glacier ice also occurs on the moraine side of the cave but exists as an ice-core beneath morainic rubble. Some of this had formed when the glacier built up against the headwall. As the firn-pack crested higher than the headwall, the glacier deposited the moraines along the top of the cliff. This zone of ice-cored debris, therefore, became easily disconnected from the main glacier during its next interval of downwasting and retreat. As the crystal sizes in this older ice are very large (some the size of a man's head) and in some sectors do not reveal a preferred fabric. They seem to suggest considerable age and growth under downwasting conditions involving a long period of stagnation and relaxation of flow stress within the ice. The fact that some of this ice is also free of air bubbles and dirt is transparent for distances up to 2 meters, suggesting that it may also relate to old, marginal lake ice from an enlarged moat-lake. It could, however, be the relict of basal melt-water frozen near the axial position of earlier tunnels. The occasional presence of a thin layer of silt at the interface of some of this old dense ice and its overlying bubbly glacier ice would seem to corroborate this interpretation.

It is concluded that this glacier tunnel remains in the same position from year to year. It probably changes little except at the cave entrances where, since 1965, drastic changes have been observed from year to year. In fact, in 1970 and 1973 a subsidiary cave entrance was found near the western entrance, as described below.

A Distributary Drainage Channel

In 1970 a secondary drainage cave entrance was found near the bottom of Lake Linda and about 10 meters directly beneath the main tunnel entrance. During 1971 and 1972 this entrance remained hidden by grounded bergs, but at the end of the summer of 1973 it had reopened enough for a team to explore and map it for a slope distance of 479 feet (145 m) in a N-NE down-glacier direction (Fig.25), as opposed to the W-SW across-glacier level of the main tunnel previously described. At its deepest point of penetration the passage thinned to a width of 1 meter and a height of 3 meters. Through this hole, a swift cold river of melt-water cascaded along the glacier bed. Down to this point, 11 survey stations were installed and distances between them measured with steel tape and compass. From its entrance this segment of the cave plunged steeply downward (attaining 41° slope), descending in a sinuous pattern for some 50 meters in the first 90 meters of its course. In this interval between stations 1 and 8 (Fig. 25), the cross-section of the tunnel was quite uniform...i.e., 2 to 3 meters wide and 2 to 3 meters high. This arching corridor was highly scalloped and composed of dense and foliated basal ice. On the floor, large unstable boulders rested, presumably extending a morainic zone from the glacier headwall. In places, series of rocks which were held fast in the ice protruded from the walls and ceiling and always appeared in straight rows. The larger dimensions of these rocks never exceeded 0.5 m. (Lamorey and Willis, 1974).

At station 6 on this transect, another passage forked back uphill toward the south (Fig. 25), along the eastern wall of which a foaming turbulent drainage stream flowed. Pieces of broken firn on the floor of this passage for some 50 meters along the junction indicated twin drainage channels from the lake.

Comparison of 1972 and 1973 Maps and Cave Characteristics

By remapping the cave in successive 12-month intervals during the two-year period 1972-73, it was found that the cave system, in spite of repeated flooding, remained essentially intact. The only substantial differences were found in the firn-pack sections of the entrances. In 1973 much of the Lake Linda entrance was covered by much more firn because of below-normal ablation in that cool spring and summer. In the extremely cool summer of 1974, the lake entrance never did open, nor did the lake drain until mid-September, after the research team had left the field.

In 1973 the character and configuration of the main tunnel remained essentially as in 1972. The drainage pattern within the cave also was largely unaltered. One noteworthy change was the appearance of the roof stalactites. A 10-meter ice column photographed and mapped in 1972 had melted to about half its length. Also missing were numerous smaller ice columns, their places having been taken by splashing water-falls from the base of moulins in the cave ceiling.

Although the overall tunnel system had not changed greatly, a more precise survey in 1973 revealed the alignment of the main cave was slightly north of that plotted in 1973. The 1973 map represents a more accurate survey because a magnetic anomaly in the cave affected Brunton compass readings in 1972 but did not alter the 1973 traverse made by theodolite.

Origin of the Cave System

The Lemon Glacier cave system appears to be related to periods of climatic amelioration in historic time...i.e., presumably to the period of warming in the 1930's and 1940's (see previous discussions on climatic periodicities, including Figs. 16 and 19). It could relate, however, to that in the 1860's (Miller, 1963, 1972b; McKenzie, 1971). The cave location may even ally to an open drainage channel over a warm interval in the 9th to 10th centuries or a lesser one in the 12th to 13th centuries (Miller, Egan and Beschel, 1968; Miller, 1967, page 213; 1975b). It may be assumed that during these periodic shrinkages of ice from the Lemon Glacier headwall a contemporaneous moat developed between present-day Lake Linda and Lynn Falls. Today this provides a significant marginal drainage way and at times of greatest deglaciation it has held a much enlarged moat-dammed lake.

In subsequent colder intervals, snow dirfted across this contact zone, keeping the channel open at least at depth. The situation is similar to what is often seen today when open fissures, such as crevasses, bridge over with drift snow, with the fracture remaining open but buried for years. There also may be sub-glacial ablation caused by water and air circulation along the ice/snowdrift contact.

The unusual character of this cave system suggests that it may represent a new category of glacier cave. We refer to such a cave as a composite glacier cave to distinguish it from those strictly produced by sub-glacial drainage or found in the lee of overriding ice masses...such as the ablation and obstruction types of Peterson and McKenzie (1968; McKenzie, 1971). The parallel moraines along the crest of the Lemon Glacier headwall are further evidence of the formerly expanded nature of ice in this headwall sector.

Of further interest is the fact that along the main cave at station 8 (91 meters in), the rock floor abruptly levelled off into a 5-meter high ice-floored room, littered with gravel. In effect this was an englacial alluvial outwash beyond which the passage floor was composed entirely of ice. Our mapping crew of Lamorey and Willis followed this section for a distance of 65 feet (20 m) farther on, at which point they found it inundated by a stream from beneath the east wall. In exploring the next 100 feet (30 m), two more gushing streams were found to enter the cave from beneath the west wall. At this farthest point of penetration, the cave rested some 80 meters beneath the névé. This englacial drainage channel probably extends all the way to the glacier terminus, some 6.5 km down-valley and every year presumably remains charged with turbulent melt-water.

As additional work is needed to delineate this unique englacial and sub-glacial drainage system, we plan to continue this research. Because of the environmental hazards involved in glacier bursts, the study has much practical value.

B. TYPICAL HYDROLOGICAL MEASUREMENTS ON THE PTARMIGAN GLACIER, 1972

Meteorological Variables and Hydrometrics

The stream gaging method used on the Ptarmigan Creek is described in Appendix D, and includes a graph of the water levels recorded. The parameters of discharge (D) of a Temperate glacier are equated as follows: $D = P + A - E$. In this, P is precipitation, A is ablation and E is evaporation.

Because of the extensive maritimity of the Lemon/Ptarmigan Glacier system, the effects of evaporation are negligible. Thus in a typical week of recorded data at the Ptarmigan Glacier stream gaging site in 1972, rainfall and ablation were the only significant factors in the liquid balance development of total propagated water. (On the Cathedral Glacier, in the continental sector of the Icefield (Fig. 2), evaporation is found to be a significant element in the liquid and mass balance...as considered later.)

When measurements of ablation on the Lemon and Ptarmigan Glacier were incomplete, we extrapolated assumed ablation from the parameters of temperature and radiation. Windspeed was taken into account, but for the 1972 summer observation period was found to be minimal. Ablation did not vary appreciably on cloudy days (v. Camp 17 correlated cloud cover, advection, temperature and ablation records over a few days in mid-July 1972, Fig. 27). Peak ablation occurred at times of high solar radiation. Peak periods of rainfall on the 4000-foot upper part of the Lemon and Ptarmigan Glaciers (Camp 17) experienced a time delay, or lag, of nearly one half day in the discharge of Ptarmigan Creek, as recorded at the gaging site at 2500-feet elevation and one mile below the Ptarmigan Glacier terminus (Fig. 8a). In contrast, peak periods of rainfall occurring directly at Camp 17A within 100 meters of the Ptarmigan Creek gaging station were followed by instantaneous rise in the stage.

The foregoing observations are substantiated by detailing some pertinent 1972 data (Fig. 28). Automatic weather station (MRI) data from Camp 17 reveal peak periods of rainfall at the top of the glacier occurred at 0700 on 14 July and at 2130 on 16 July, 1972. Following these highs were stream level peaks at 1800 on 14 July and 0930 on 17 July. These times are each to within a half hour of actual times and therefore precipitation on the upper névé is indeed followed by stage-discharge at the terminus in about 11 hours.

In contrast, by using the three-hourly synoptic meteorological records from the lower valley station (Camp 17A), it is seen that peak rainfall periods included the times on July 15, 1972 from 0300 to 0700 and from 1000 to 1300 (Fig. 29). Peaks in the water stage recorder at the 17A stream gage site (Fig. 28) were recorded on 15 July at 0400 and again at 1030, indicating that for Ptarmigan Creek almost no time is required for an increase in discharge following the onset of significant precipitation in the immediately adjacent valley.

On 18 July, 1972, no precipitation was recorded at any station. Therefore, the only new source of discharge at the end of this day had to be by ablation. The ablation peak occurred at the time of peak temperature and peak radiation... a condition which took place between 1300 and 1430 (compare Figs. 27 and 28). It is of interest, then, that the peak in the Ptarmigan Creek stage recorder site occurred on July 18 at 2000, indicating a time lag between peak ablation and discharge of 4-1/2 to 7 hours. If a 5-1/2 hour lag is involved, this agrees closely with the expected value of discharge from mid-glacier.

The discharge records from the Ptarmigan Glacier during the cited observation interval (July 12 to 19, 1972) are given in Figure 28, followed by Figure 30 which extends the daily discharge record for Ptarmigan Creek through Sept. 16. In Figure 27, the precipitation recorded by the MRI unit at Camp 17 over these same days is noted, with the data rendered in hundredths of inches per hour. Note again that precipitation at Camp 17A, taken in the 3-hourly synoptic increments, is also noted for comparison in Figure 29. Ablation records on the upper Lemon Glacier (Camp 17) over this observing period are noted in Figure 27.

The first peak of the discharge curve on 12 July 1972 is related to the ablation peak on that day (Figs. 27 and 28). The remaining discharge peaks (as previously discussed) show direct correlation with rainfall. In a subsequent report, these data will be analyzed with respect to the U.S. Geological Survey Lemon Creek Stage records to see whether in this field season there were anomalous peaks in the Lemon Creek outflow introduced by the presence of Jokulhlaups.

For further evaluation, in Figure 27 (upper) data on temperature, radiation, and cloud cover at Camp 17 are plotted for this 8-day observing period, together with records of relative humidity and dew point. Additional analyses of these records and others obtained during the summers of 1971-74 are beyond the scope of this report.

Ptarmigan Glacier Névé Surface Measurements

In addition to continuous recording devices employed for temperature, humidity, precipitation, wind, and radiation, meteorological readings were taken by station personnel every three hours. A party hiked each evening from Camp 17 to Camp 17A and returned in the morning, permitting continuous collection of diurnal data at each station. From these, the effects of weather parameters operative at each station were noted.

The Theissen Polygon Method was employed in the analysis. The névé basin was divided into five areas and nine appropriate ablation observing sites were positioned, one or two to represent each area. The sectors and key observation site are noted in Figures 31 and 32 as Camp 17 (4200'), site Betty (3700'), site Chris (3200'), site Joan (2700') and Camp 17A. On the schematic the approximate "basin" drainage areas in square meters are indicated to permit unit area calculations.

From the recordings of rainfall at Camps 17 and 17A and from the radiation records at Camp 17, it was possible graphically to represent where and when ablation and rainfall occurred. With such factors known, it was anticipated that we could determine the effects of these controls on the ablation and discharge from Ptarmigan Creek at the glacier's outlet.

Surprisingly the weather on the Ptarmigan Glacier was found to differ over its entire length, largely due to strong orographic influences...i.e., slope and confining side-walls. In the five polygonal sectors, rainfall was observed to fall at different times and in unequal amounts. Because of this and the long time between data collection at the three on-ice stations, only the precipitation and ablation data at these points were obtained. A week of these data, covering the same time span as the other cited records, is presented in Figure 33. The histograms show the considerable differences in precipitation and ablation pertaining at the three mid-glacier sites. Fuller details of the hydrological parameters involved in this research are given in the following section.

C. HYDROLOGICAL COMPLICATIONS AND MEASUREMENTS IN THE PTARMIGAN DRAINAGE BASIN*

A detailed report of the 1972 investigations of the hydrological characteristics of the Ptarmigan Glacier and its highly maritime drainage basin is presented here. The outflow stream, Ptarmigan Creek, is tributary to Lemon Creek, in the valley below and to the north of the Ptarmigan Glacier drainage basin. The ongoing study of hydrological aspects of the watersheds of the Lemon, Ptarmigan and Thomas Glaciers, being conduted by Maynard Miller and Austin Helmers, has previously been reported on (Miller, 1972b).

Purposes of the Study

The objective of this continuing study is to observe the relationships between various causal factors and the discharge of the glacier via Ptarmigan Creek. The approach is to consider the environment, including terrain relationships, and to spell out the most important causal factors and to experiment with methods of describing them. Some of the observations and the calculations based upon them are, of course, approximations, but in the analysis the relative importance of various factors can be recognized. The experience gained in the field served to inspire new channels of inquiry and to provide further insights to help plan and execute further field studies in this glacier system and in the comparison study of the Cathedral Glacier on the continental flank of the icefield.

During the 1972 summer season, the Stevens Hydrological Stage Recorder (previously noted as set at Camp 17A) was used to provide continuous records of water levels in Ptarmigan Creek. Numerous stream gagings during the summer served as the basis for translating the data on stream stage into discharge data (Appendix D). Detailed meteorological records were kept at locations within the Ptarmigan drainage basin. In fact, seven temporary observation sites were set up in addition to the two permanent Camps 17 and 17A to determine how various parameters varied throughout the drainage basin. Their relative positions on a longitudinal profile of Ptarmigan Glacier are noted in Figure 32. Thus it is possible to consider in more detail what effects solar radiation, air temperature and precipitation have upon stream discharge.

Data Acquisition Relating to Stream Flow

The Stevens Recorder at Camp 17A maintained continuous records of the level of Ptarmigan Creek from late June into September, 1972. The stage records used in this study are synthesized in Figures 28 and 30, each segment of which represents one day of record. Because discharge (in cubic feet per second) is more useful than stream stage for many calculations, it was essential to correlate stream stage and discharge in accordance with the method noted in Appendix D. A reference wire was stretched across the creek slightly upstream of the Stevens Recorder. During each run, the stream was divided essentially into fifteen trapezoids of two-foot base and height equal to the water depth. This was facilitated by the markings on the wire. The velocity of the water at the center of each trapezoid was

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* Prepared by Lawrence Brush, formerly of the Geology Department, Temple University and currently of the Geology Department, Harvard University, and Maynard M. Miller, Geology Department, Michigan State University and the Foundation for Glacier and Environmental Research.

then measured with a rotating water-cup gage and multiplied by the area of each trapezoid to give the discharge through each element of area. These were summed over all areas to give the total stream discharge. This figure was then plotted against the mean stage recorded on the Stevens Gage during the time interval (about one-half hour) in which the gaging was done. Each run required about three man-hours for gaging and calculations. This fact, along with the necessity to make runs over as wide a range of stream stages as possible, dictated t t the stream gagings be performed over a number of weeks. Two groups performed the gagings. The first group performed 12 gagings and the second 20, for a total of 32 gagings. These groups of data are considered separately and together in Figures 34 and 35. It is interesting to note the good fit of these data (even when considered separately) to the same straight line. Thus conditions which determine the slope of these plots evidently did not change over the data collection period. It is assumed that these conditions did not change over the entire summer season when the plots obtained are used to interpret stream stage data from times other than during this particular data collection period (which covered the month of July). Deviation from the straight line is greatest for the lowest values of stage and discharge. This is because the gage used to measure velocities approached the lower limit of its accuracy in the shallow water associated with discharges of less than 25 cubic feet per second. Limits of this type do not apply to the Stevens Stage Recorder in this study. Inspection of Figures 34 or 35 allows one to determine the discharge in cubic feet per second represented by any point on the plots in Figures 28 and 30.

An independent experiment was carried out by L. Brush to determine the temperature of stream waters at the point at which they flowed out from the fields of old snow still retained below the terminus of the glacier and to see how quickly the temperature of the waters rose as they flowed away from the glacier. The temperature was measured in the fastest flowing, deepest waters every 100 meters along the main creek bed. During each run, several readings were made at each point and the average taken. It was assumed that water flowed off the snowfields below the terminus at 32°F. The average temperature recorded there was slightly different from 32°F, probably due to the calibration of the thermometer, and was thus set equal to 32°F. Other readings were adjusted accordingly. Despite the fact that the equipment available was imprecise for this task, a temperature gradient along the Ptarmigan Creek was detected (Brush,'74, with upwards of a 3.3°F rise in water temperature within a half mile of the snowfields through which the water flowed. The measurements were made at times of maximum discharge (late evening).

Configuration of Drainage Basin

The Ptarmigan drainage basin is located near the southeastern end of the Juneau Ice-field, on its southwestern (maritime) margin. The axis of the drainage basin runs north-south. The southern (upper) end of the drainage basin is located about five miles from the city of Juneau. The drainage basin lies in the map area of the U.S. Geological Survey Juneau B-1 Quadrangle (Fig. 23).

The longitudinal profile of the Ptarmigan drainage basin (Fig. 32) was constructed using an enlargement of the quadrangle cited above. Locations of the two permanent research stations used in this study, Camps 17 and 17A as well as the locations of the seven temporary field sites, are noted in Figure 32 . The curve ABC in the profile represents the central axis of the drainage basin. The curve BD is that part of the trail between Camps 17 and 17A which departs significantly from the axis of the drainage basin. Thus, Camp 17 and two temporary sites (Kathy and Betty) were located off the center line axis of the drainage basin.

The drainage area was calculated using an enlargement of the aforementioned quadrangle. It was obtained by summing the areas of simple geometric figures constructed within the boundaries of the basin (Fig. 31). This sum represents the area of the horizontal surface enclosed by a line formed by the intersection of a horizontal plane and the right cylinder generated by the projection of the boundaries of the basin parallel to a vertical axis. The area was deemed appropriate for calculations of volumes of precipitation* and was calculated to be about 1.66 square miles (or the equivalent in square meters as noted in Figure 31).

The areas of the drainage basin which are and are not covered by glacial ice are hereafter referred to as presently glaciated (glacierized) and deglaciated areas (de-glacierized). The average elevation above sea level of the drainage basin and the average elevations of the glaciated and deglaciated areas of the drainage basin respectively were calculated as follows. A square grid was superimposed on the enlargement of the topographic map. The elevations of the points within the drainage basin that were located by the points of the grid were noted and averaged. It was also noted whether or not these points were located on the Ptarmigan Glacier. Because the accuracy of such a determination is directly proportional to the number of points that are counted, a square grid was constructed with spacings small enough to locate 289 points within the drainage basin. On this basis it was calculated that 46.7% (0.76 square miles) of the Ptarmigan drainage basin was covered by the glacier at the time that the surveys for the map were made (1948); 53.3% (0.88 square miles) were deglaciated at that time. The average elevation above sea level of the drainage basin as a whole was calculated to be 3450 feet. The average elevations of the presently glaciated and deglaciated areas were calculated as 3600 feet and 3300 feet respectively. All determinations of elevation in these calculations were made to the nearest fifty feet.

The fact that the survey for the topographic map was conducted in the late 1940's means that the drainage basin profile (Fig. 32) and the figures for areas and elevations cited above are off somewhat, but at least this limitation is known. Downwasting of the Ptarmigan Glacier during the last twenty years or so is responsible (Fig. 8a). Thus the area of the drainage basin covered by the glacier and the average elevation of the glaciated areas are slightly overstated. Similarly the average elevation of deglaciated terrain (3300 feet) is probably understated. But variations in the amount of terrain covered by ice and snow that occur as melting proceeds during the summer season are probably more significant than the former variations. Thus behavior of the drainage basin with respect to runoff caused by melting of ice and snow varies as well during the course of the season. The figures cited above, although approximations, show that the hydrological characteristics of the drainage

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* Another area could be calculated. It would represent the area of the atmosphere-drainage basin surface interface. If dA is an element of area as calculated above (the area of a horizontal plane) and θ is the angle of dip of the terrain beneath the element of area, then this other area, A', can be defined by setting $dA' = \frac{dA}{\cos\theta}$. Because of the steepness of the terrain under consideration, $A' = \frac{dA}{\cos\theta}$ would be larger than $A = dA = 1.66$ square miles. A' might prove useful in calculations involving evaporation, sublimation, or solar radiation.

basin cannot be completely due to effects of the glacier. Relatively large areas (in terms of percentages of total area) which are deglaciated also contribute to the runoff. These deglaciated areas are mainly to the north of the terminus.

Because of inaccuracies caused by glacial retreat, the average elevations of the glaciated and deglaciated areas of the drainage basin will not be considered further at this time. Only the average elevation of the drainage basin as a whole is cited in correcting precipitation and temperature observations made at a single research station to a normative value for the drainage basin as a whole.

Effects of Air Temperature and Solar Radiation on Stream Discharge

The synoptic meteorological records maintained at Camps 17 and 17A during the summer make it possible to relate air temperature and solar radiation to the hydrological characteristics of the Ptarmigan drainage basin. Based on micro-meteorological studies on the Lemon Glacier (v. Wendler and Streten, 1969), radiation and temperature influence stream discharge by controlling the rate at which snow, firn and permanent ice melt. The relationship between radiation and temperature and stream discharge, however, is complicated by several other factors including that proportion of runoff caused by precipitation which must be recognized and separated from the effects caused by radiation and temperature. Another complication arises because radiation and temperature may influence stream discharge in ways other than by direct control of the rate at which frozen water melts. It is assumed, however, that with the intense maritimity of this locale, evaporation is slight and hence radiation and temperature influence stream discharge essentially by controlling rates of melting.

Solar radiation may also be the most important causal factor at times when there is no precipitation...i.e., via ablation. Unfortunately, the relationship between radiation and discharge cannot be described quantitatively here. Thus the effects of temperature, which are amenable to such treatment at this time, are considered in detail (v. Figs. 27, 36 and 37).

The air temperature at a given point also varies with time. The temperature also varies significantly from place to place within the drainage basin. The latter type of variation poses the problem of choosing which temperature to relate to stream discharge. The solution is approached here by thinking in terms of a mean temperature of the drainage basin.

Elevation Effects on Ambient Temperature

Because of the topographic relief, elevation is assumed to be the major cause of temperature variations from place to place at a given time. The change of temperature with respect to elevation (i.e., the local lapse rate) is considered here as the vertical temperature gradient of the drainage basin. The temperature gradient is defined as the partial derivative of temperature with respect to vertical distance (elevation). It is a partial derivative because the temperature of a point varies also with time. Field data can be used to determine the vertical temperature gradient of the drainage basin and to develop a means of determining the mean temperature of the drainage basin, given the temperature at a single research station.

Simultaneous meteorological observations were made at Camps 17 and 17A from 0400 July 13 to 2200 July 18 and from 0100 July 25 to 1900 August 6, 1972. During these

periods, observations were made every three hours for a total of 286 simultaneous temperature observations. During these 19 days of simultaneous observations (Figs. 27, 35 and 37), temperatures at Camp 17 (elev. 4185') averaged 3.1°F. below those at Camp 17A (elev. c. 2400'). Two temperature inversions in the records which occurred within these temperature periods were averaged into the above figure.* The average temperature difference cited above is simply a mean. It averages temperature differences which vary in accordance with changing weather conditions. The validity of applying a single-value representation of the temperature difference to all meteorological situations (in summer) is probably commensurate with the approximation that elevation alone is the cause of temperature variations within the drainage basin at a given time.

Other Temperature Controlling Factors

Some factors besides elevation which influence temperature differences within the drainage basin can be mentioned. There is an interesting relationship between the temperature difference between Camps 17 and 17A and the time of day. For example, the temperature differences were greater in the afternoons than at other times (Fig. 37), at least during the period of simultaneous observations. Wind speed and direction, cloud cover (especially at night), and the intensity of solar radiation probably influence the temperature difference between places with different elevations. The period of simultanaous observations at Camps 17 and 17A was too short to allow consideration of the relationships between these factors and the temperature differences within the drainage basin, with the exception of the relationship between temperature difference and time of day.

The mean temperature difference of 3.1°F. between Camps 17 and 17A (based on the noted 286 simultaneous observations over 19 days) can be used to calculate a temperature gradient for the drainage basin and to derive an expression for the temperature at any point as a function of the temperature at a single research station. To do this, one must use the elevations of Camps 17 and 17A and make the assumption that temperature is a linear function of elevation. Because meteorological records are much more extensive for Camp 17 than they are for Camp 17A, it is more convenient to consider an expression which gives the temperature at any point in terms of the temperature at Camp 17. This expression can be written as

$T = T_{17} + \frac{3.1Y}{1785}$, or $T = T_{17} + 0.00174Y$, where T is the temperature at the point

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* A possible cause of the first inversion illustrates the variability in solar radiation from place to place within the drainage basin, which is one reason that radiation connot be related quantitatively to stream discharge. A thick cloud cover blanketing the area for the whole day of July 17th broke up about 1600 (Fig. 27). The consequent increase in solar radiation that presumably caused Camp 17's temperatures to rise did not take place at Camp 17A because mountains just to the west of the lower part of the valley effectively shaded this camp from the sun at that time of day. No explanation is offered for the second inversion.

in question; T_{17} is the observed temperature at Camp 17; 3.1 is the mean observed temperature difference between Camps 17 and 17A; 1785 is the elevation difference in feet between Camps 17 and 17A; and Y is the difference in elevation between Camp 17 and the point in question. This expression for the temperature of a point as a function of the Camp 17 temperature can be modified to give the average temperature of the drainage basin as a function of the Camp 17 temperature. The average elevation of the drainage basin was shown previously to be about 3450 feet. Thus the average value of Y becomes equal to 4185 - 3450, or 735, and the expression for the average temperature of the drainage basin becomes $T = T_{17} + 1.28$. This expression is used hereafter to determine the mean temperature of the drainage basin for periods during which observations were made only at Camp 17.

During periods of simultaneous observations the mean temperature is determined by interpolation to the value of 3450 feet. It is the mean temperature of the whole drainage basin that is used to correlate temperature and stream discharge. The temperature gradient of the drainage basin is the slope of the family of straight lines generated by inserting different values of Camp 17 temperatures into any of the above expressions and graphing the results. Here $\partial T/\partial x$ is equal to 0.00174°F. per foot. Other expressions for the temperature at a given point as a function of the Camp 17 temperature and for the temperature gradient of the drainage basin could be derived by using the relationship between temperature difference and time of day (Fig. 37). But these variations in temperature difference between Camps 17 and 17A were neglected to simplify consideration of the relationship between temperature and stream discharge.

Examination of Stream Stage Records Re Flow Maxima and Lag Effects

Stream flow which is influenced mainly by daily variations in solar radiation and air temperature can be distinguished easily from that which is influenced by precipitation. Examination of the Stevens Stage Recorder data (both for Ptarmigan and Lemon Creeks) is all that is needed. Stream stage versus time curves which show the dominant influence of radiation and temperature are characteristically smooth with rounded extrema. Typical examples are shown in Miller, 1972b; in Jones, 1975, and Bugh, 1973 as well as in Figure 30 of this report.

All daily stream stage plots versus time that did not appear to represent the effects of precipitation to the exclusion of radiation and temperature effects are also shown in Table VIII It can be seen that stream discharge was maximum in the late afternoon and evening between 0850 and 1600. The mean time of the stream discharge peaks was 2212 hours. The mean time of the stream discharge minima was 1148 hours. The highest temperatures on the days in question were recorded at the 1000, 1300, 1600, and 1900 observation times while the lowest temperatures were recorded at the 0100 and 0400 observation times. The lags between the observation times at which the highest temperatures were recorded and the times of the maximum peaks in stream discharge ranged from 1 to 13-1/2 hours. The lags between the observations which recorded the lowest temperatures and the times of the minima in stream discharge ranged from 9 to 13 hours. Less variability is shown by the latter lag times because the low number of 0100 and 0400 meteorological observations precluded calculation of more than five such lags. These figures are not based on thermograph records but are valid enough to show that daily extrema in air temperature have little effect on daily extrema in stream discharge. This view is reinforced by the wide scattering of points that obtains when the minimum temperatures are plotted against the minima in discharge.

TABLE VIII

CHARACTERISTICS OF PTARMIGAN CREEK FLOW INFLUENCED BY RADIATION AND TEMPERATURE

DATE 1972	TEMPERATURE				WATER STAGE						DISCHARGE (CFS)		RADIATION			PRECIPITATION (in)	
	MAX	TIME	MIN	TIME	MAX H20 LVL	TIME	LAG BEHIND MAX TEMP (HRS)	MIN H20 LVL	TIME	LAG BEHIND MIN TEMP (HRS)	MAX	MIN	INDEX	TIME OF MAX	HRS SUN	24 HRS PRIOR	24-48 HR PRIOR
19 July	52.9	1130	45.9	0400	.97	0100	13.5	---	----	----------	48	---	2.50	1300	10	0	0
20	----	----	----	----	1.08	2330	----	.89	1030	----------	56	40	----	----	--	-	-
21	----	----	----	----	1.06	2300	----	.95	1130	----------	55	45	----	----	--	-	-
22	----	----	----	----	.97	2330	----	.89	1400	----------	48	40	----	----	--	-	-
23	----	----	----	----	.93	2100	----	.86	1300	----------	44	37	----	----	--	-	-
24	----	----	----	----	.87	2200	----	.82	1400	----------	38	33	----	----	--	-	-
25	52.3	1300	44.7	0400	.83	0130	12.5	.79	1600	12	33	30	2.09	1300	1	.07	0.
26	43.6	1900	40.4	0100	---	----	----	.81	0830	7.5	--	32	1.31	1600	0	.29	.07
30	50.6	1300	38.6	0400	.70	2200	9	.69	1400	10	23	21	3.04	1300	9	0	.22
31	56.6	1430	42.4	0100	.76	2400	9.5	.64	1400	13	28	18	2.75	1300	13	0	0
1 Aug.	55.9	1600	45.6	0400	.79	2230	6.5	.66	1300	9	30	20	2.87	1300	8	0	0
10	53.5	1900	42.2	----	.95	1830	----	.88	1200	-----------	45	39	----	----	--	0	.46
11	56.3	1600	45.3	----	.90	1930	3.5	.80	1000	-----------	41	31	3.07	1300	15	0	0
12	48.5	1600	43.3	----	.81	1900	3	.78	1100	-----------	32	29	1.84	1300	0	0	0
17	56.3	1300	46.3	----	1.04	1600	3	1.02	0830	-----------	53	52	1.80	1300	1	.02	.69
18	53.3	1600	45.3	----	.97	1830	2.5	.86	1100	-----------	48	37	3.07	1300	8	0	.02
19	57.3	1000	47.3	----	.93	1830	8.5	.77	1000	-----------	44	28	2.68	1300	12	0	0
20	55.3	1600	46.5	----	.94	2000	4	.76	1030	-----------	45	28	1.99	1600	8	0	0
21	56.1	1600	47.8	----	1.01	1930	3.5	.84	1030	-----------	52	35	2.65	1600	10	0	0

Sunshine, Radiation and Runoff

A general lack of correlation between temperature parameters and the indices of stream discharge is noted in Table VIII. Some days noted in this table display anomalous results because precipitation occurred during these days or on prior days. (For example, on July 25, 26, and 30, and August 10, 17, and 18.) Removal of these days from consideration, however, does not improve the correlation significantly. It is doubtful whether the use of thermograph records would clarify these discrepancies. It is believed that they arise because air temperature is a factor of much less importance than solar radiation in controlling melting and runoff. Further, the effects of precipitation may extend several days beyond the actual period of rainfall. Changes in the rate of percolation of water through the firn-pack and ice and in the glacial drift of the lower valley also influence the processes and lag effects in runoff due to melting of frozen water (Leighton, 1952).

It is unfortunate that the effects of radiation could not be more quantitatively related to runoff, although Wendler has given some ratios with respect to firn melt (op. cit.). The radiation index data in Table VIII are merely a sum of the radiation intensity readings taken at each meteorological observation at Camp 17. The duration of sunshine figures are averages (when possible) of data from Camps 17 and 17A and are more often the results of observations made at Camp 17. The effects of shading by mountains, the strike of the terrain, and weather patterns make it necessary to adjust results such as those which are taken at one or two stations before they can be used to interpret the behavior of these variables throughout the basin as a whole (v. Andrews previous discussion). Such a treatment focused at this research locale will permit better correlation to be made between radiation, temperature and discharge. Such is also essential to any further heat budget studies of the drainage basin. The fact that such a treatment of air temperature (i.e., via establishment of the concept of a mean temperature of the drainage basin based upon observed differences in temperature data from Camps 17 and 17A) did not yield a good correlation of air temperature and stream discharge leads to the conclusion that other factors, notably radiation, exert the major effect on runoff from this glacier.

Sample Computations

Two calculations illustrate some aspects of the problem of determining the effect of radiation and temperature on the melting of snow and ice. August 12, 1972 was a completely overcast day, with temperatures averaging in the mid 40's. It can be seen with the aid of Figure 27 and Figures 30a,b,c that about 2.47×10^6 cubic feet of water flowed out of the drainage basin via the Ptarmigan Creek in a 24-hour interval from 1100 August 12 to 1100 August 13. This roughly corresponds to the time interval between two discharge minima. If one makes the assumption that all this runoff was derived by direct melting of ice and snow, one can immediately calculate the energy required to cause the melting which resulted in this runoff. Taking a value of 80 calories for the energy required to melt a gram of solid water (which is in turn responsible for one cubic centimeter of runoff), we can calculate a requirement of $(2.47 \times 10^6\ ft^3) \times 2.8 \times 10^4\ \frac{cc}{ft^3}$

$\times 1\ \frac{gram}{cc} \times 80\ \frac{cal}{gram} = 5.52 \times 10^{12}$ calories. Assuming that by August 12th, one square mile of the drainage basin was covered by ice and snow, we can speculate as to how much area of ice and snow cover was needed to provide a unit volume of

stream discharge. In this one square mile is equal to about 2.48×10^{10} cm^2. The amount of heat absorbed per unit area that was used to melt the ice.. which appeared as runoff was therefore $\frac{5.52 \times 10^{12} \text{ cal}}{2.48 \times 10^{10} \text{ cm}^2} = 223 \text{ cal cm}^{-2}$. Since about 80 calories are required to melt a gram of frozen water, the amount of runoff produced by a square centimeter of ice and snow surface was $\frac{223 \text{ cal cm}^{-2}}{80 \text{ cal gm}^{-1}} = 2.79 \frac{\text{gm}}{\text{cm}^2}$. This is equivalent to 2.79 cubic centimeters of stream discharge from each square centimeter of ice, firn or snow surface exposed in the basin.

Another example: August 18th was partly cloudy with temperatures that averaged (Fig. 27) around 50°F. On this day, about 3.38×10^6 cubic feet of water flowed out of the drainage basin via the Ptarmigan Creek between 1100 August 18 and 1100 August 19. Assuming again that all this discharge was the result of the melting of frozen water that day (after correction for the lags between extrema in radiation, temperature and stream discharge is made) and if one notes that 3.38×10^6 cubic feet of water are equivalent to 9.49×10^{10} grams of water, then the amount of energy required to cause this melting was 9.49×10^{10} gm x 80 cal gm^{-1} = 7.58×10^{12} calories. The number of calories per unit area of snow and ice surface that caused melting is

$$\frac{7.58 \times 10^{12} \text{cal}}{2.48 \times 10^{10} \text{ cm}^2} = 306 \text{ cal cm}^{-2}.$$

Thus each square centimeter of snow and ice surface supplied about 3.82 cubic centimeters of the stream's discharge.

The difference between these two sets of figures may illustrate the difference between a day with total cloud cover and temperatures in the mid-forties (August 12) and a day with partly cloudy skies and temperatures around 50°F. The only assumption in the above calculations that was checked was the unstated assumption that the creek water just below the terminus of the glacier was at a temperature of 32°F. A more rigorous study of the effects of radiation (and the resulting complication of temperature effects) on runoff would make possible a detailed study of the heat interactions between radiation, ice and snow, and runoff (v. Pinchak, Part II-J of this report).

Effects of Precipitation on Stream Discharge

Detailed records were also kept at several research sites in the Ptarmigan drainage basin to examine the relationships between precipitation and stream discharge. During the study it was determined that the amount of precipitation varies from place to place within the drainage basin. Two objectives motivated investigation of this phenomenon. The first was to describe this variation, while the second was to devise a basis for formulating estimates of the volume of precipitation that falls within the area of the drainage basin based on field data.

To describe the variations in precipitation, meteorological records from research sites within the drainage basin were compared. The two permanent stations, Camps 17 and 17A, provided the most reliable records. Early in the summer, two temporary

stations, designated Betty and Chris, were established on the Ptarmigan Glacier. At the same time, a third station, Joan, was set up just north of the terminus. Later in the season four more temporary sites were established (Barb, Irene, Jennie and Kathy). The locations of these stations are shown in Figure 32.

During the month of July, it was found that precipitation varied regularly along the north-south axis of the drainage basin and that the higher (southern) stations generally recorded more rainfall than the lower (northern) stations. It is considered that precipitation variations along this axis are either (1) a function of elevation; (2) a function of horizontal distance along this or some other axis; ~~(3) a function of~~ both elevation and distance, or (4) a function of relief variations in the terrain. To describe precipitation variations with accuracy , it was essential to maintain simultaneous records for as long as possible. Several problems arose. For example, only four standard rain gauges were availabe. These were deployed at Camps 17 and 17A, and at sites Chris and Joan. At the other four stations measurements were made with improvised rain gauge equipment. The improvisations worked well but in a few cases there were interruptions in the simultaneous data collection. Thus the comparison periods appear staggered. Human error ruled out the use of some of the data collected at Camp 17.

The periods of data collection at Camps 17 and 17A constitute the longest interval of simultaneous observation. Records were made from 2200 July 12 to 2200 July 18, and from 2200 July 24 to 2200 August 4, 1972. During these periods, the total precipitation at Camp 17 was 8.24 inches and that at Camp 17A was 6.15 inches. The precipitation at Camp 17 exceeded that at Camp 17A by a factor of 1.34. These ~~results are graphed (Fig. 3) with precipitation plotted against elevation~~. Precipitation is again assumed to be a linear function of elevation. Taking into account the average elevation of the drainage basin (calculated as 3450 ft.), one can use this plot to estimate the mean amount of precipitation which fell within the drainage basin during these periods. Multiplication of this figure by the total area of the drainage basin gives a value for the volume of precipitation during these periods. The period of data collection was too short to allow this treatment to be extended to periods for which the records of only one station are available. In comparing precipitation and runoff, records of Camps 17 and 17A were used. Again, precipitation is assumed to be a linear function of elevation. This allows interpolation to a figure for the precipitation which fell at 3450 feet. This normative value then allows calculation of the volume of rain that fell within the entire drainage basin during any interval of simultaneous observations at Camps 17 and 17A.

Although calculations of precipitation volumes are based only on the records from Camps 17 and 17A, it is interesting to note the results of observations made at the seven temporary glacier sites. Figures 40a and 40b show the results of the precipitation profile from data collected at five of these temporary sites from 1600 July 27 to 1300 August 8, 1972. This 12-day period covered 50 per cent of a storm which lasted from 1600 July 25 to 1600 July 29 and 100 per cent of a storm which lasted from 1600 August 2 to 1300 August 9. Figure 40a shows the precipitation at each station plotted against the elevation of each station. Figure 40b shows precipitation plotted against distance along the axis of the drainage basin. Although stations Betty and Kathy were not located directly on this axis their precipitation totals may also be plotted after allowances are made for this deviation. The dashed lines on both graphs represent extrapolations of precipitation quantity to the elevation and positions of Camps 17 and 17A. Precipitation values for these two permanent stations could not be plotted on this graph for the interval 1600 July

27 to 1300 August 8 because of some interruptions and also the existance of some questionable data. The extrapolation for Figure 40a predicts that Camp 17 precipitation would have exceeded Camp 17A's precipitation by a factor of 2.37. The extrapolation for Figure 40b predicts that Camp 17's precipitation would have exceeded Camp 17A's precipitation by a factor of 2.86 during the same period. The difference between 2.37 and 2.86 and the observed factor of 1.34 for the periods 2200 July 12 to 2200 July 18 and 2200 July 24 to 2200 August 4 is noted. This difference may in part be a function of the brevity of the periods of simultaneous collection of data.

Simultaneous observations of precipitation at Camp 17A and five of the temporary sites over the period 1600 July 27 to 1500 August 6 covered 50 per cent of the storm which lasted from 1600 July 25 to 1600 July 29 and 58 per cent of the storm that lasted from 1600 August 2 to 1300 August 9. Camp 17 data for this particular period could not be used because of inconsistencies. The results of these observations are shown in Figures 41a and 41b which respectively plot precipitation against elevation of the stations and distance along the drainage basin axis. An extrapolation to the elevation of Camp 17 in Figure 41a predicts that precipitation there would have exceeded that at Camp 17A by a factor of 2.97. An extrapolation to the distance between Camps 17 and 17A in Figure 41b predicts that Camp 17 precipitation would have exceeded that at Camp 17A by a factor of 2.59.

Camp 17 precipitation records coexist with those of five of the temporary glacier sites for the period 1600 July 27 to 1500 August 4 and from 1500 August 6 to 1300 August 10. These periods cover 50 per cent of the storm from 1600 July 25 to 1600 July 29 and 43 per cent of the storm from 1600 August 2 to 1300 August 9. The amounts of precipitation are plotted against elevation in Figure 41a and against geographical position in Figure 41b. Extrapolations to the elevation and geographical position of Camp 17A yield factors for the amount of rainfall at Camp 17 relative to that at Camp 17A of 1.06 and 1.08 respectively.

Comparative Melt-Water and Rainfall Precipitation Analysis

Results of the comparison of Camps 17 and 17A over the period 2200 July 12 to 2200 July 18 and 2200 July 24 to 2200 August 4 (Fig. 39) are thought to give the most realistic representation of variations in precipitation. Firstly, the time interval covered is the longest. Also the two stations are the nearest to the upper and lower extremes of the basin. Furthermore, the equipment used at these two stations was the most reliable. A factor of 1.34 for the relative amounts of rainfall at Camps 17 and 17A is thus held to be the most realistic. It is recognized, however, that 17 days is too short a period of data collection to permit generalization of the results to the whole summer season.

Stream flow that is influenced by the runoff caused by rainfall precipitation is easily distinguished from that caused by the runoff of melting ice and snow. This has been well reported in previous JIRP reports (Miller, 1954). Flow which is strongly influenced by rainfall precipitation shows sharper extrema than that caused by melting ice and snow. Precipitation extrema in stream flow do not generally or necessarily occur at the late evening and mid-morning times characteristic of flow which is not under the influence of rainfall precipitation. Changes in flow caused by variations in solar radiation and air temperature can sometimes be discerned during periods of rain precipitation. Also if ambient temperatures are at or about 40° F., precipitation must itself cause some melting of ice and snow (Miller

1972b). If 100,000 cubic feet, or 2.83×10^9 cubic centimeters, of rain at 40°F (4.5°C) were to fall and interact with snow and ice so as to cause the maximum amount of melting possible, there would be enough heat to melt 1.57×10^8, or 5570 cubic feet of water, from the snow and ice mass. This figure is equal to about 5.5 per cent of the rainfall. This assumes, again, that water flows off the glacier at about 32°F.

Two storm periods were studied in some detail. These periods were from 1600 July 25 to 1300 July 29 and from 1600 August 2 to 1000 August 9, 1972. These were the only precipitation periods throughout the summer when simultaneous meteorological observations were made at Camps 17 and 17A every three hours around the clock. For these intervals, rates of rainfall and stream discharge and temperature were compared. Over these periods, there were no hydrological surges recorded on the Stevens Recorder graphs. Discharge rates in cubic feet per hour were thus computed only for the times which coincided with meteorological observations (0100, 0400, etc.).

Discharge rates for a given time were computed as follows. Stream stage for a given time was read off the Stevens Recorder graphs (Figs. 28,30:Ptarm. Cr.). By means of the results of stream gagings shown in Figures 34 and 35, these water depths were converted to discharge rates in cubic feet per second. The figures were then multiplied by 3600 to give discharge rates in cubic feet per hour. Rainfall rates were computed in terms of the number of cubic feet per hour which fell on the drainage basin. Precipitation measurements, made at Camps 17 and 17A, which covered three-hour data periods were noted. Based on these figures the average rainfall for the drainage basin as a whole for the three-hour interval was determined by interpolation. The elevations of Camps 17 and 17A and the average elevation of the drainage basin are 4200 feet, 2500 feet and 3450 feet respectively. To interpolate, again, one makes the assumption that precipitation is a linear function of elevation and that it determines the amount of rain that would have been recorded at 3450 feet during the three-hour interval. To get an hourly rate in terms of cubic feet, we multiply the above figure by one-third and then by the area of the drainage basin. If we designate the volume of rain that falls within the drainage basin (in cubic feet) as Y and the average (3450') value (in inches) of rain as X, then $Y = \frac{1}{3} \times \frac{1 \text{ foot}}{12 \text{ in.}} \times X \text{ inches} \times 4.57 \times 10^7 \text{ feet}^2$ or $\underline{Y = 1.27 \times 10^6 \text{ cubic feet per hour.}}$

By graphing this function (Fig.42) rapid determinations are possible. The temperatures used in the following comparisons are average temperatures of the drainage basin as a whole. They are calculated by the same sort of interpolation to the 3450 foot level as used above for precipitation data, using Camps 17 and 17A meteorological records.

These parameters are plotted for the two above mentioned precipitation periods in Figure 38 a, b and c.

The precipitation period 1600 July 25 to 1300 July 29, 1972 (Fig.38a) appears to show the effects of both daily solar radiation and precipitation in the stream discharge rates. Increases in hourly rates of stream discharge show smooth increases through the afternoons and evenings of July 25, 26, and 27. These increases resemble the reactions of the Ptarmigan Creek to daily solar radiation. But on August 12, a cloudy day with average temperatures in the mid-forties, the highest runoff rate (at 1800) was only 119,000 cubic feet per hour. The peaks of July 26 and 27 far

exceed that, although on these days temperatures averaged in the low to mid-forties. Further, the increases in the hourly rate of stream discharge on July 25, 26 and 27 follow closely peaks in rainfall centered about 2030 July 25, 1430 July 26 and 0830 July 27. For these reasons it is concluded that the increases in stream discharge are due to the combined effects of radiation and precipitation.

July 28, 29 and 30 show decreasing rates of stream discharge as average temperatures rose (solar radiation was approximately constant). This effect marks the lower amounts of precipitation that fell on these days. It is interesting to note that on July 28, 29 and 30, the rate of stream discharge fell during the early morning hours. On days when there was no precipitation, discharge also fell during the early morning hours in delayed response to the cessation of solar radiation at sunset and falling air temperatures. It can be concluded from this period of precipitation that as long as the rate of rainfall does not exceed the maximum rate of stream discharge caused only by melting ice and snow (due mainly to solar radiation) the record of stream discharge will reflect to an equal degree peaks in solar radiation and precipitation.

Figures 38b and 38c show the same parameters for the precipitation period 1600 August 3 to 1300 August 10, 1972. During this period, rates of rainfall exceeded those shown in Figure 38a by factors of as much as 4.0. Stream discharge rates exceeded those shown in Figure 38a by as much as a factor of 4.65. On at least five occasions the rate of rainfall exceeded the rate of stream discharge. Periods for which data are missing are indicated with dashed lines in Figures 38b and 38c. (These dashed lines do not infer any type of behavior and are inserted for convenience only.) Peaks in rainfall rates, centered about 2030 August 2, 1730 August 3, 0530 August 4, 1130 August 4, 2030 August 4 and 0830 August 8, correlate with responses in stream discharge. Peaks in stream discharge, centered about 2200 August 5, 0400 August 6, and 1000 August 8, presumably indicate that concommitant gaps in precipitation data were times of peaks in rainfall rates.

After the rainfall ended, stream discharge declined (on the average) for a few days even with higher average temperatures on August 10, 11 and 12. This lag, also seen after the precipitation period shown in Figure 38a, indicates the existence of a delay mechanism of some sort. Further evidence for such a mechanism is seen on August 3 and 4, 1972, when rainfall rates far exceeded runoff rates. The areas under these curves represent volumes of water. It can be seen that the three areas above the discharge curve and below the rainfall curve can be fitted into the areas over which discharge exceeds rainfall. Delays on the order of several hours are the only way to explain the excess of rainfall over discharge. Possible examples of such mechanisms might be the seepage of water through glacial drift and the impounding of water for periods of time in crevasses and ice dammed (englacial) pools. If the effects of solar radiation were amenable to more quantitative description than has been accomplished on this glacier system to date, it might be possible to describe the behavior of these delay mechanisms in more detail.

Concluding Comment

Several of the relationships between meteorological causal factors and stream discharge have been examined. The runoff from melting snow, firn and ice caused by solar radiation and air temperatures above freezing and by the influence of rainfall precipitation has been discussed. It is hoped that this evaluation adds to the conclusions

developed in reports on previous field seasons of research on the Lemon and Ptarmigan Glaciers (e.g., Andress, 1962; Bugh, 1966; Zenone, Helmers and Miller, 1967; Miller, 1969, 1972b; Zenone, 1972). Further steps towards understanding the nature of the complications in the glacier's liquid balance resulting from these observed effects of terrain configuration await a more rigorous examination of the role of solar radiation (Wendler and Streten, 1969) in inducing the melt process in glacier firn and ice and, from this, in causing runoff.

D. GLACIO-HYDROLOGY OF THE LEMON-PTARMIGAN GLACIER SYSTEM DURING THE 1973 ABLATION SEASON- A YEAR OF DELAYED FIRN MELT*

The water-ice balance of the Lemon-Ptarmigan glacier system in the heart of the area between the main cellular circulation of the continental anticyclone and the cyclonic North Pacific Low is representative of the glaciers in the Alaskan Panhandle and has been under hydrologic surveillance since 1965. These studies are of added interest because of emphasis on global glacio-hydrological research during the International Hydrological Decade (1966-74).

The Lemon Glacier and its adjoining Ptarmigan Glacier have also been the focus of general glaciological research by the Juneau Icefield Research Program since 1949. Currently, a continuing hydrological and related mass balance study is underway on these glaciers. They have been chosen because of their accessibility (6.5 km northeast of Juneau), their simple configuration and unity, and their advantageous position with respect to a U.S. Geological Survey gaging station on the Lemon Creek, where runoff records have been maintained consecutively since 1951. Additional advantages are the proximity of the U.S. Weather Bureau (ESSA) stations at the Juneau Airport, Juneau City and Annex Creek, and the summer research program weather stations on the adjacent Juneau Icefield, as discussed in previous sections of this report. Some of the 1973 glacio-hydrological investigations on this prototype glacier system are noted in the following section.

Glacio-Meteorology, 1973

For the summer field season of 1973, synoptic records were begun on June 16 at Camp 17 on the Lemon-Ptarmigan Glaciers, both thermo-physically Temperate ice masses on the southern flank of the Juneau Icefield. Jones in a previous chapter has shown that under average weather conditions consistent differences were not observed between Camps 17 and 10, the latter essentially at the same elevation but in the southern interior of the icefield. This indicates that the icefield's south central sector is also significantly influenced by maritime conditions and, conversely, that Camp 17, nearest the ocean, is influenced by continentality to about the same measure at least as the nearer inland stations within 25 miles. Thompson in his previous chapter, however, has shown that at least for stormy periods Camp 17 is the most maritime in character. It has also been shown by Miller (1972b) that the maximum precipitation occurs in the southern sector of the icefield (even more so than at Juneau) but despite onshore flow throughout stormy periods the precipitation amounts are quite uniform across the icefield. Past experience has shown that daily temperature variations are for the most part clearly parallel, although of course not equal, at all stations, suggesting that competent records obtained even over short periods at a single field station can have regional significance.

Weather conditions observed during a period from June 16 to July 23, 1973 are tabulated in Appendix D. Notable is the cloudiness and precipitation which must be considered atypical for this part of Alaska in summer (Little, 1972a). The low interdiurnal variation in temperature and the high relative humidity are indicative of the maritime nature of the region, but are not what would be expected

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In addition to ARO-D support, this investigation was supported by grant No. 023-7167A, Research Foundation of State University of New York.

in the mean at this location when compared with the long-term mean data for summer at Juneau (Little, op. cit). The association of the mean daily ablation with the daily mean values of cloudiness, temperature and relative humidity are evident in Figure 27, as considered in preceding section.

The low diurnal range of temperature and humidity further indicate the maritime nature of the location. Over this observation period, cloudiness showed no marked diurnal variation, but was possibly slightly higher in the late afternoon. Diurnal variations h ablation were not determined in 1973.

The period of analyzed observations is relatively short, but it appears that the climatic pattern cannot be considered typical of the long-term mean conditions over the early ablation season in this region. However, if these conditions represent a secular fluctuation of climate and mass balance trends and are considered in terms of a shifting position of the Arctic Front, the present conditions reveal the inland shift of storm tracks and the phasing in of more maritime conditions on the Juneau Icefield...an affirmation of Miller's conclusions in recent years (1963, 1972b, 1973).

Weather and the General Circulation of June 1973

A well-developed Gulf of Alaska Low was a prominent feature of the 10,000-foot (700 mb) height map for June 1973 (Fig. 43). This Low was slightly east of its May position (Dickson, 1973), as were the other major components of the wave train over the Pacific Ocean (Taubensee, 1973), in response, partially, to the continued strength of the mid-latitude westerly flow. Wind speed along the axis of maximum 700-mb wind from south of Japan to the Washington-Oregon coast averaged at least 5 meter/second faster than normal over the entire route. The maximum departure, in excess of 11 meters/second, was reached in the strong height gradient between the Gulf of Alaska Low and the broad subtropical ridge to the south (Taubensee, op cit).

Mean 700-mb height departures fell over much of western North America from May to June as short-wave impulses from the mean Low in the Gulf of Alaska moved eastward across the continent. The blocking High over western Canada weakened slightly and moved northward while the southern portion of the associated ridge was less amplified than it had been in May (Taubensee, op cit).

These circulation relationships for the months of October 1972 to October 1973 are noted in the diagrams in Figure 43.

Snow/Firn Surface Conditions and Seasonal Terrain Changes

The snow cover, or firn-cover, may be continuous, with a smooth, gently, undulating or slightly rippled surface, or it may be broken up by active or stabilized drifts (or by suncups in summer). On the Lemon and Ptarmigan Glaciers, the snow surface exhibits a patterned microrelief. The surface features may initially be caused by accumulation of snow or by erosion. The long axis of the pattern found in snow appers to parallel the wind, but the uniformity of the pattern is usually quite striking. Future research will determine the relationship of the duration and intensity of the prevailing winds to the amplitude and frequency of snow surface features.

When the ablation season starts, the snow surface becomes pock-marked, or rilled, as meltwater moves through the pack, and eventually the surface exhibits ablation

features as sun cups (Little,'72a).Thus it is almost impossible to apply a rational classification scheme to the patterns and features which produce characteristic snow and firn scapes. However, since snow-surface or firn-surface conditions in different seasons may control operations in regions dominated by snow or firn cover, it is essential that some information about such surface features be included in snow and firn observation reports.

In 1969, Dr. Edward Little of the Juneau Icefield Research Program initiated a terrain roughness project to create artificial smoothing of suncups. The 1969 ablation study was only of limited success on the Lemon Glacier. Dr. Little continued these studies on the Camp 10 névé in 1970, at Camp 25 in 1971 and at Camp 29 in 1972 (some aspects of which are briefly noted elsewhere Little, 1972a). He concluded that the large quantities of chemical (hexadeconal) necessary to accomplish effective smoothing over large areas render the method impractical. For special micro-surface research and in other small area applications, however, chemical spreading may find a unique use. Bugh (1972) described the possibility of forest fire control of climate, using an illustration taken from the 1969 ablation season. In this season smoke from forest fires filled the skies for many days and much of the smoke settled on the Lemon Glacier, resulting in a relatively smooth surface

Surface Conditions of the Early 1973 Ablation Season

Through the middle of June, 1973, the surface had few suncups or similar topographic features. Later in the month, small rills were evident with an accumulation of dust. The rills were generally oriented northeast-southwest. Suncups were present on June 27-28 but were less than 2 cm deep. Rain through June 29-30 almost completely destroyed the suncups but the rills persisted. The rills can be traced to cols suggesting wind origin.

Suncups were again present on the upper névé on July 3 and aligned with the wind similar to transverse dunes. The surface was pocked by July 10, but rain began on the 10th, washing out most of the suncups by July 13. The rills were still very prominent nonetheless.

Ptarmigan Glacier Surface Characteristics

Rills were accentuated by dust and red algae by June 26 but no suncups were present on the upper névé. Rill orientation was northwest-southeast. By July 1, the rills attained a depth of 5 cm. Suncups were present on July 4 on the upper névé. The lower névé was deeply pocked (15-25 cm) with suncups accentuated by red algae and dust accumulation. Rills of 30 cm depth were very common on the lower névé on July 4.

Firn-Pack Hydrology

In 1965 (Bugh, 1966), a new research station was constructed on the ridge between the Ptarmigan Glacier and the Lemon Glacier to serve as a base for continuing study during the upcoming International Hydrological Decade (IHD). The Lemon Glacier constitutes what may be called a representative basin for research on hydrological systems as a basic component of the IHD program. The hydrological records and previous glaciological work show that the Lemon Glacier has been gradually diminishing in size over the past four decades. Bugh (1966, 1972) and others (Miller, 1972b) have shown that this trend is reversing itself.

Continued research is up-dating, amplifying and analyzing glacio-hydrological observations from 1948 to 1975. The water/ice balance is being investigated with respect to runoff and retained volumes of ice in each glacier, adjusted for effects of impounding, capillary retention and precipitation trends... aspects of which have been dealt with in a number of the studies reported on in this present report.

Records of englacial temperatures have been made during each ablation season, and liquid water movement in the firn-pack measured with collecting funnels installed at selected pit levels. Deployment of these funnels reveals that usually after the end of August downward percolation of propagated surface water is confined almost exclusively to the upper 125 centimeters.

Meltwater movement has been studied by means of specially designed galvanized iron funnels which were connected by a rubber tube to a collecting bottle. These funnels were placed in niches excavated in the north- and west-facing walls of the pits to a horizontal depth of approximately one meter and then walled-in with firn so that they could not be influenced by radiation. Such meltwater movement studies have been conducted on the Taku Névé by Leighton (1952) and Miller (1954). Additional studies were conducted on the Lemon Glacier in the latter part of the 1965 ablation season. The results show large amounts of water are continually percolating through the firn and that the actual magnitude of flow is dependent on meteorological conditions (Bugh, 1966).

The funnels were placed in the firn-pack to determine what influence was exerted on water movement by bands of ice. The results show that meltwater percolates to a depth slightly greater than one meter (August 18-2?, 1965). The zero degree Centigrade isotherm was then at a depth of 90 cm where an ice band was forming. Ram penetrometer profiles show that the ice band was continuous throughout the Lemon Glacier and prevented percolation below its depth. Therefore, within the névé of the Lemon Glacier much of the surface propagated meltwater was recaptured at a depth of less than one meter.

Meltwater movement was also studied in the Lemon Glacier firn-pack during the early 1969 ablation season. Meltwater refroze only a few centimeters below the surface where an ice band was observed increasing in thickness until July 8 when intercepted by the ablation surface.

Englacial Temperature Measurements and their Regime Implications

The early ablation season of 1973 was characterized by a normal pattern of surface melting. However, collecting funnels deployed in the firn pack received almost no propagated water. This can be explained, in part, by the facts that (1) the National Weather Service at Juneau recorded a departure of -6.38°C for the water balance year to the end of June and (2) by the firn-pack's low temperature of -5.0°C on June 22, 1973. As late as July 17, 1973, the temperature of the firn-pack was -1.0°C, which was most unusual for this Temperate maritime glacier. These low temperatures measured in 1973 may have hindered the movement of meltwater on the Lemon and Ptarmigan Glaciers but few extensive ice layers resulted from the accompanying refreezing of meltwater.

Ram penetrometer profiles (Bugh,'74) do not indicate that any significant ice bands formed in the firn-pack during the interval June 19 through July 17, 1973. The lack of ice band formation accompanied by the low firn-pack temperatures could indicate that in this abnormally cool summer the névé surface lowering was

accomplished in part by evaporation and compaction rather than melting and infiltration, although evaporation is not normally found in this strongly maritime locale.

Collecting funnels were placed in the firn-pack under rills (Fig. 44a) on July 21, 1973. On July 22, one funnel had collected 3510 ml of water (over a 20 hour,45 minute-period).On July 23, an additional 3850 ml accumulated (over an 18 hour, 45 minute-period). Surface lowering would appear to be accomplished through meltwater movement along rill zones. The rills appear similar to stream channels (Fig. 44a, b). Further investigation could parallel the methods employed by fluvial geomorphologists.

Surface "Drainage" Patterns, Their Geometry and Possible Explanation

Stream channels draining the land are integrated into varied drainage patterns. Drainage networks have highly probable patterns when analyzed statistically. On a gently sloping surface of homogeneous rocks, a random pattern forms, with the possibility of stream flow in all directions being equal. A steeply sloping surface will cause a deviation from the ideal distribution of stream-flow direction.

Morisawa (1968) discusses the "laws" of drainage composition and shows deviations suggesting that the "laws" simply represent statistical relationships. The hydraulic geometry of the rills on the Lemon and Ptarmigan Glaciers deviate from the "laws" of drainage composition to such an extent that the basic nature of the rills must be different from that of streams (except for the larger runnels).

Orientation of rills and runnels (Fig. 44) suggests that they are of wind origin (or sub-surface water flow on buried ice strata...Miller) rather than surface water drainage channels. The fact that some rills rise over topographic surfaces (Bugh, 1974) suggests that they are not of running water origin. Following their inception, many rills serve as traps of algae and dust. The dark color of these substances reduces the local albedo and accentuates melting, hence deepening the rills.

Surface lowering on the upper nèvè of the early ablation season (June 20 to July 22, 1973) averaged approximately 4 cm per day. Meltwater movement through the firn-pack was essentially non-existent during this time. Again, evaporation and compaction in this unusually cool summer would seem to be a better explanation than melting and infiltration (usually associated with warm summers) as the cause of surface lowering.

Percolation funnels collected considerable quantities of water after July 21, when set in place under rills of the upper nèvè on the Lemon Glacier, although no meltwater movement was observed in the rills. This suggests that the rills are not channels by which meltwater moves but that the July 21 date is more significant in the ablation history. The firn-pack was below 0°C through July 17 and little water was noted in any percolation funnel to that date. On July 21, however, the firn-pack became essentially isothermal,i.e. at 0°C (see below) and significant quantities of water were collected immediately following this date.

Firn-Pack Temperatures at Given Depths- Lemon Glacier Nèvè (4100')

22 June '73	30 cm	50 cm	100 cm	150 cm	200 cm	300 cm
	0°C	0°C	-1.0°C	0°C	-1.0°C	-1 0°C
21 July '73	0°C	0°C	0°C	0°C	0°C	0°C

E. DETECTION OF THAWING FIRN IN HYDROLOGICAL BALANCE DETERMINATIONS THROUGH NEAR-INFRARED IMAGERY FROM ERTS-I SATELLITE *

In the summer of 1972, NASA launched its Earth Resources Technology Satellite (ERTS) which has a sun synchronous, near-polar, near-circular orbit. Looking down from 567 miles, it views areas every 18 days. The size of the synoptic view and the ground resolution of ERTS instruments dictate that experiments in hydrology cover regional targets, such as large lakes, soil-moisture distribution, snowfields and icefields.

Earlier work has shown that hydrological balance studies can be aided by determination of the seasonal névé-line position (data since 1945 given in Tables IX and X). Remote sensing by ERTS of allied shifts in transient snow-lines and seasonal névé-lines could enable the plotting of short-time interval changes in dimension and position. With the addition of such continually monitored information, greater precision can be added to the determination of annual mass and liquid balances of glaciers.

Table IX Late Summer Névé-Lines and Hydrologic Balance of the Juneau Icefield (Data from Miller, 1963, 1972b, 1975d; Egan, 1965)

Re[illegible]e Year	Seasonal Névé-line Elevation (feet) in the Taku Glacier Sector	General Hydrologic Balance of Coastal Icefield Sector
1974-75	2600	Strongly Positive
1973-74	2800	Positive
1972-73	2850	Positive
1971-72	2900	Equilibrium
1970-71	2800	Positive
1969-70	2700 ±	Equilibrium
1968-69	2400 ±	Strongly Positive
1967-68	2900 ±	Equilibrium
1966-67	3050	Near Equilibrium
1965-66	3100	Near Equilibrium
1964-65	2650	Strongly Positive
1963-64	2450	Strongly Positive
1962-63	2850	Positive
1961-62	2950	Equilibrium
1960-61	2900	Equilibrium
1959-60	3100	Equilibrium
1958-59	3000	Equilibrium
1957-58	3050	Negative
1956-57	3200	Negative
1955-56	3200	Negative
1954-55	2550	Surplus
1953-54	3200	Negative
1952-53	3300	Negative
1951-52	3100	Equilibrium
1950-51	3430	Strongly Negative
1949-50	3300	Negative
1948-49	2600	Strongly Positive
1947-48	2850	Positive
1946-47	2950 +	Equilibrium
1945-46	3200 ±	Negative

* From Dr. James E. Bugh, Dept. of Geology, State University, Cortland, N.Y.

Table X Hydrological Balances, Lemon Glacier, 1945-75 *

Budget Year	Seasonal Névé-Line (Elev. in ft.)	Hydrological Balance
1974-75	3300	Slightly Positive
1973-74	3550	Negative
1972-73	3650	Negative
1971-72	3700	Negative
1970-71	3650	Negative
1969-70	3500	Negative
1968-69	3250	Positive
1967-68	3500	Negative
1966-67	3700	Negative
1965-66	3600	Negative
1964-65	3200	Positive
1963-64	2900	Strongly positive
1962-63	3150	Strongly positive
1961-62	3650	Slightly negative
1960-61	3550	Slightly negative
1959-60	3700	Negative
1958-59	3750	Negative
1957-58	3450	Negative
1956-57	3225	Negative
1955-56	3600	Negative
1954-55	2625	Positive
1953-54	3500	Negative
1952-53	3375	Negative
1951-52	3225	Negative
1950-51	3450	Negative
1949-50	3375	Negative
1948-49	3150	Positive
1947-48	3300	Negative
1946-47	3350 (?)	Negative
1945-46	3150	Positive

*1948-58 data from Marcus, 1964; other data from Miller, 1963, 1972b, personal com.

The Multi-Spectral Scanner (MSS) of the satellite is a line-scanning device using an oscillation mirror to scan terrain. Four synchronous images are produced, each at a different wave band. The wave-lengths for the bands used in this study are: Band 5 (lower red)...0.6 to 0.7 micrometers and Band 7 (near IR)...0.8 to 1.1 micrometers. Band 7 is best for land-water discrimination while Band 5 is best for topographic and cultural features, such as drainage patterns, roads and towns.

The Juneau Icefield appears very black in the Multi-Spectral Scanner Band 5 imagery (Fig. 45a) of 11 August 1972. Its appearance is distinctly different in Band 7 imagery (Fig. 45b) of the same date. In Figure 45b, the terminal areas of the Taku, Norris, Lemon, Mendenhall, Herbert and Eagle Glaciers show a brightness reversal (light in tone).

Fresh snow appears light in tone on both Bands 5 and 7 (dark in Fig. 45), because the prints were made from 70 mm black and white transparencies. The near-IR brightness reversal of the glacier termini is due to melting of the firn-pack and absorption in the infrared while still reflecting in the visible. Strong et al (1971) in a review of the literature on spectral albedo show fresh snow under clear skies to display almost constant reflectivity over the range of the visible spectrum. Reflectance of wet snow in the visible is invariant with wave length but slightly less than that for dry snow. However, a 16 % drop in reflectance for wet snow has been observed in the near-IR compared with an 8 % drop for dry snow.

The extinction coefficient of pure water for wave lengths of band 7 (0.8 to 1.1 micrometers) is approximately 1 (Strong, et al, op. cit.). For a beam of incoming radiation to penetrate a surface water film, reflect from an underlying ice or snow surface, and pass again through the water film, the water film must be considerably less than 1 cm thick. Considerable absorption would be expected for even a 1 mm layer of water. Thus the appreciable reflectance drop in the near-IR would not require a visible layer of w[illegible]

Meteorological data for Juneau on August 11, [illegible] mean temperature of 59°F (15°C). This substantiates melting conditions on t[illegible] lo[illegible] ts of the icefield. The following listing shows the elevations of the nea[illegible] htness reversal for this 11 August 1972 imagery on six key glaciers.

Icefield Elevations of Near Infrared Br[illegible] Reversal, 11 Aug. 72

Glacier	Elevation (ft.)
Taku	2600
Norris	2200-2600
Lemon	3000
Mendenhall	2500
Herbert	3100
Eagle	2700

On the lower firn-pack, the ablating firn absorbed a sufficient amount of incident radiation to cause it to appear light in Figure 45b. At higher elevations, the firn-field remained frozen and highly reflective in the near-infrared.

The extent of ablation on the Juneau Icefield, as suggested by Figure 45 implies a transient névé-line position for each of the six glaciers listed above.

Although August 11 was not significantly late in the ablation season, the possible névé-line positions do suggest a probable mass and liquid balance for the 1971-72 hydrological year. Névé-lines of 2600 feet and 3000 feet for the Taku and Lemon Glaciers respectively indicate equilibrium balances for both glaciers (see previous tabulations of hydrological balance, Table IX, re mid-September positions).

Whereas the results of this study are preliminary and the explanation only tentative, a case is made for the detection of névé-lines on the glaciers of the Juneau Icefield and on other icefields around the globe by remote sensing from earth satellites. This detection method takes advantage of the strong absorption of incident radiation in the near-infrared (Band 7) by a covering film of water. The two-band measurement, of course, provides further information on firn and ice conditions beyond that of merely delineating the boundary of ice and snow-covered areas.

F. COMPARATIVE LIQUID WATER CONTENT OF THE TAKU, LLEWELLYN AND CATHEDRAL GLACIER FIRN-PACK*

Firn Stratigraphy, Upper Taku Glacier Névé, August, 1972

For comparison with similar profiles in previous and subsequent years of the Juneau Icefield Research Program, test-pit measurements were made on the upper Taku Névé at elevations of 5,850 feet (1725 m) near the Camp 8/18 Junction and at 6,100 feet (1850 m) near the Camp 25/26 Junction on the crestal névé of the Taku-Llewellyn transection glacier system. Only representative data are here discussed. These were obtained between late afternoon of August 19th and early morning of August 20th, 1972. Stratigraphy, density, grain size and free-water content determinations were made. The resulting data are shown in Figure 46.

(1) Stratigraphy and Grain Form

As shown in Figure 46, the late-summer stratigraphic thickness of the 1971-72 firn-pack on the main high-level Taku Névé was 15'1" (4.51 m). It was 13'7" (4.1 m) in the pit at the crestal névé, at an elevation 250 feet (76 m) higher. This latter value represents total retained firn above the 1971 annual summer ablation surface just south of the "sun-line" on the upper Llewellyn Glacier.

In the firn of each profile there were many bubbly ice plates (strata) and ice lenses into which melt-water from solar radiation and rain had been refrozen. The firn grain sizes averaged 1-4 mm in diameter and were rounded in form, having experienced the first stage of constructive metamorphism. The larger crystals were found immediately above a thick ice stratum delineating the 1971 late-summer surface. This zone was comprised of depth-hoar crystals, as manifest by their quite coarse texture. Some of the crystal shapes were prisms and pyramids and some were cup-shaped...i.e., hollow forms, in spite of free-water contents of 3.8 to 7.0 per cent. This means that the depth-hoar was initiated after the end of the 1970-71 summer ablation season, probably well into late September. The base of the annual accumulation increment was quite easily identified by this distinctive zone and the annual ablation surface (A.S.,Fig. 46) it represents.

Summation: The annual accumulation for 1970-71 on the 5850-foot (1725 m) névé near the Camp 8/18 Junction of the upper most Taku Glacier was 8.27 feet (2.506 m or 2506.5 mm/cm^2) in water equivalent. At the crestal site on the uppermost Llewellyn Glacier Névé it was slightly less...i.e., 7.97 feet (2.415 m or 2415.0 mm/cm^2) water equivalent. Respectively, this represented a mean of approximately 15 feet (4.6 m) of late summer firn.

(2) Density Profile

As shown in Figure 46, the density profile at the Camp 8/18 Junction was fairly uniform at every level, except above the 0.6 m depth. In the crestal névé profile, the density gradually increased from the firn surface to about one meter

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* From field measurements by C.W. Kreitler, Dept. of Geology, University of Texas and F. Nishio, University of Hokkaido, Japan, with the assistance of the 1972 Glaciological Institute participants.

depth. Below this depth, it remained uniform at about 0.58 g/cm^3 to 3 meters and was variable down to the 1971 ablation surface. This suggests that the bulk density of firn at the Camp 8/18 Junction was 0.557 g/cm^3 and at the test pit near the Camp 25/26 crestal névé junction was 0.575 g/cm^3.

Free Water Content

For liquid water content measurements the calorimetric method was employed using the equipment and procedures noted in Appendix F. The free water content at the Camp 8/18 Junction pit gradually decreased from 21.1 per cent on the snow surface to 3.8 per cent at a depth of 14 feet (4.24 m), as noted in Figure 46. The free water content in the crestal névé was stable at about 6.6 per cent to 10.6 per cent, except it was 0 per cent at a depth of 0.6 m, probably reflecting a refreezing in the cold air of the middle of the night when these data were obtained.

An acceptable mean for the free water content values is approximately 8 per cent of the amount of annual accumulation which comports with the mid-summer averages of 12 to 30 per cent in the firn-pack at the elevation of Camp 10 (i.e., near 3600 feet (1090 m) as determined in previous JIRP seasons, Miller, 1954, 1972b).

In Section G additional firn stratigraphy and free water content measurements are reported in the 1972 firn-pack...one at a comparable elevation and three at lower elevations on the Taku Glacier. It is noted that additional firn profiles were obtained at these sites in 1971, 1973 and 1974, but the presentation of all these comprehensive records is beyond the limits of this report.

Free Water Content Measurements on the Cathedral Glacier Névé

On the upper névé of the Cathedral Glacier, hourly water content was measured with the calorimeter during daylight hours of September 7, 1972. The test-pit location was above the névé-line at an elevation of 5850 feet (1770 m). (Fig. 47) This equates to the mean elevation of this highest névé. At the same time, meteorological data were obtained at Camp 29 (Fig. 48), as well as surface stage records from the glacier's pro-glacial lake at 5600 feet (1700 m) elevation. Stream level records in the outlet stream down-valley from the terminal moraine complex were obtained during the summers of 1972 through 1974. Representative data from these records are discussed in the following section.

In 1972, the Cathedral Glacier had a very small accumulation zone (névé), indicating a quite negative mass balance for that year. In 1973 and 1974 the annual mass balance regimes were also negative, but less so (contrasting Taku ' I,Table IX)

Test pit measurements at the same place as the free water content measurements are graphed in Figure 49. One significant observation from these data is that the firn (density about 0.560 g/cm^3) quite suddenly changed to bubbly ice at a depth of 130 cm. This suggests that firn becomes glacial ice at shallow depths on this glacier and that snow and firn are transformed to ice in as short a period as two to three years, probably via the process of refreezing of melt-water into superimposed ice.

The same kind of stratigraphy has been recognized in the accumulation area on McCall Glacier, located north of the Arctic Circle in the Brooks Range, Alaska... a glacier which is classified as sub-Polar. Englacial temperature measurements

on the Cathedral Glacier in 1972 also suggest that this glacier is thermophysically sub-Polar.

Free water content was measured at different depths below the snow and firn surface, i.e., at 20 cm, 50 cm and 100 cm (Fig. 47). On the firn surface, the FWC seemed to have increased with solar radiation shortly after noon and to reach a maximum value of about 13 per cent whether the firn surface was in the shade of the cirque headwall or not. (When the surface was in the shade, a frozen surface extended to a depth of about 7 cm). At depths of 20 and 50 cm, the FWC was uniform at about 5 per cent, but at a depth of 100 cm it showed fluctuations influenced apparently by percolating water.

As shown in Figure 48, the relation between ambient air temperature and FWC on the firn surface was striking. Strong reductions in FWC were established within one hour after the nèvè became shadowed. Good correlation was also found between FWC reduction on the firn surface and variations in lake level. This showed up within 2 to 3 hours. For example, it took this amount of time for run-off water melted by solar radiation from the firn surface to flow down and reach the proglacial lake.

Also, as expected, run-off water levels in the outflow stream located below this lake might be correlated directly with lake level stages, but the stream stage appeared to experience peaks of flow one to two hours earlier than the water level peaks on the lake stage recorder (Fig. 48). This is a seeming paradox which relates to other factors considered in Section I below.

G. FIRN DENSIFICATION AT DIFFERENT ELEVATIONS AND CLIMATES ON THE JUNEAU ICEFIELD*

During the summer of 1972, observations and measurements were made on snow and firn conditions at four selected locations at four different elevations on the Juneau Icefield representing different climatic situations. Our purpose was to consider formation of new and old snow into firn and firn into glacier ice in the temperate conditions which affect the maritime névés of the icefield.

Observation sites were chosen on the climatological transect from Camp 10 to the Camp 8/18 sector of the Taku Glacier's highest névé (or from 3500 feet (1060 m) to above 6000 feet (1820 m) elevation, v. Fig. 1). The locations were adjacent to key camps. Near Camp 10, the site was 10B, at 3500 feet elevation (1060 m) at a point one-half mile south of the field station. The Camp 9 site was at the Camp 9/18 junction at 4900 feet (1500 m) and site 8G, below Camp 8, was at 5850 feet (1725 m). On this transect, the climatic conditions gradually change from a typically maritime and geophysically temperate situation at Camp 10 to sub-temperate thermophysical conditions in a sub-continental environment in the Camp 8/18 sector, and a sub-Polar condition on the highest névés up to 7100-8000 feet (2150-2425 m) near Camp 25 on the Mt. Nesselrode/Mt. Bressler Plateau (Fig. 2). The Mt. Nesselrode section of this transect is to be discussed in a later report.

Borings were made in each location and core samples obtained from the bore-holes. Measurements were recorded on stratigraphy, density, hardness, grain size, free water content and temperature of the snow and firn in situ. The results obtained are tabulated in Appendix G and graphed in Figures 50 a and b.

The results strongly suggest that melt-water percolating into the snow and firn materially weakens the mechanical strength of snow and firn and therefore plays a significant role in the densification process. Surprisingly, densification of snow and firn seems to proceed faster in the colder regions of the icefield than in the warmer and more temperate sectors. This is well shown in Figures 50 a and b, which reveal generally higher densities and increased zones of straticulated ice and iced firn (Miller, 1955) in the more elevated test-pit and bore-hole sites on the icefield. In turn, this suggests that percolating melt-water may play a more important role in the densification process of snow in the higher and colder elevations in the accumulation zones of all temperate and sub-temperate glaciers. This probably relates to the polythermal nature of such multiple névés (i.e., where there are persistently colder englacial conditions at higher elevation).

Related glacio-hydrologic investigations were also made on the lower reaches of the Mendenhall Glacier, one of the thermophysically temperate valley glaciers stemming from an intermediate elevation névé of the icefield. This related study is described in the following section as it relates to the role melt-water plays in the growth of ice crystals and in the related process of glacier flow.

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* This research was conducted under the direction of Dr. Gorow Wakahama of the Institute of Low Temperature Science, University of Hokkaido, Japan, assisted by several graduate students in the Glaciological Institute.

H. MELT-WATER PERMEATION BELOW THE NÉVÉ-LINE *

During the 1971 summer field season, some melt-water studies were conducted below the névé-line on the Mendenhall Glacier by Dr. Gorow Wakahama and students. As this study forms such a useful transition in the interrelation between névé ablation and ultimate runoff from sub-glacial channels (noted in the preceding sections), it is included here with Dr. Wakahama's kind permission as an important adjunct to the discussions in the preceding two sections.

Melt-water permeating through a glacier body has been widely discussed in connection with recent studies of glacier surges. At the beginning of March, 1968, Dr. Wakahama and T. Takahashi found that on the Mendenhall Glacier melt-water permeated the glacier body even in early spring (Wakahama, 1969; Takahashi and Wakahama, 1970). The mean air temperature during the period of observation was a few degrees above zero Centigrade. Melt-water percolated from the wall of a bore-hole drilled to 7 m depth at the lower part of the glacier. The ascending speed, $\underline{V}$, of the water level in the hole was measured at 50 cm/hr at a depth of 450 cm and 40 cm/hr at a depth of 250 centimeters. Assuming that Kirkham's formula for the hydraulic conductivity of soil holds in this case, the coefficient of permeability of the glacier ice mass was 6×10^{-6} cm/sec, which corresponds to that of compact silt.

In the summer of 1971, in further pursuit of this study on the Juneau Icefield, the glacier body was again studied on the Mendenhall Glacier by similar methods. A network of water channels, through which the melt-water permeated, was observed both on the Mendenhall Glacier and on the McCall Glacier in the Brooks Range of northern Alaska.

Method for Obtaining Water Permeability of Glacier Bodies

The method of obtaining the water permeability of the Mendenhall Glacier was the same in principle as that for obtaining the hydraulic conductivity of soil developed by Kirkham (Kirkham, 1954). Drilling was carried out to a depth of 2-3 meters to take out ice cores using a SIPRE-type hand auger. The bore-hole was used as a well to obtain the water permeability of the glacier body. The level of water in the bore-hole was first measured just after the water stopped rising. Then, the water was bailed out of the hole with a stainless steel cylinder (50 cm long and 6 cm in diameter) which had a valve at the bottom. A small amount of salt was put into the hole to make the water electrolytic. Then a pair of vinyl wires, with an exposed junction at the tip, was hung down into the hole. At the moment when the tester connected to the wires indicated a closed circuit, the length of the hanging wire was measured, yielding the water level at a given time, and hence, the rate of rise.

Hydrometric Results

On the lower Mendenhall Glacier, about 1.5 km from the terminus (Appendix H), borings were made at three selected sites: (1) on a flat and wide area (site B1); (2) near a supraglacial pond in a concave area (site B2) and (3) on the top of a glacier ridge (site B4). The results obtained at Sites B1 and B4 are shown respectively in Figures 51a and b, in which the water level in the

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* From notes of Dr. G. Wakahama, Institute of Low Temperature Science, University of Hokkaido, Japan. Dr. Wakahama is a research affiliate of the Juneau Icefield Research Program.

hole is plotted against time. At the beginning of the measurements, the water level rose rapidly, but gradually slowed down and then saturated the surrounding ice. The final water level was observed at a depth of 10 to 80 cm beneath the glacier surface.

Assuming that Kirkham's formula from soil science can be applied to this case, the coefficient of permeability, $\underline{K}$, of the glacier body was calculated in each area:

$$\underline{K} = 0.617\ (\underline{r}/\underline{s}\underline{d}).\ (\Delta \underline{h}/\Delta \underline{t})$$

where $\Delta \underline{h}$ is the increment of the water level within $\Delta \underline{t}$ seconds, $\underline{r}$ is the radius of the hole, $\underline{d}$ is the depth of the hole from the final water level. $\underline{s}$ is a coefficient which is determined from the value of $\underline{r}$, $\underline{h}$ and $\underline{d}$.

The calculated value of $\underline{K}$ is 1×10^{-3} cm/sec in the flat and wide area; $3-5 \times 10^{-3}$ cm/sec in the concave area; and $2-3 \times 10^{-4}$ cm/sec on the glacier ridge. These values of water permeability correspond to those of fine sand.

The water permeability, $\underline{K}$, of the glacier body at the same spot varied with time. For example, at site B1, $\underline{K}$ was found to be 1×10^{-3} cm/sec on July 31, but decreased to 5×10^{-4} cm/sec on August 1. The final water level was found at the same depth in the hole on both days. The meteorological conditions on the glacier were not very different on these two days. This suggests that the situation of the water-channels in the glacier body may be changed with time. Eventually, some water-channels may become narrowed and even closed.

The water level in the glacier body was also found to vary with time. For instance, it gradually ascended toward evening as shown in Figure 51c. The water level may be dependent on the melting rate at the glacier surface. Further observations are necessary to provide more detail on the daily variation of water levels.

<u>Grain-Boundary Water-Channels in the Lower Mendenhall Glacier</u>

Although the formation of water-channels in the main ice body of this Temperate glacier has recently been discussed in conjunction with glacier flow and the question of glacier surges, very few observations on the very small water-channels have been made <u>in situ.</u>

On the thermophysically Temperate Mendenhall Glacier , as well as on the thermophysically sub-Polar McCall Glacier in the Brooks Range, comparative observations are available on the network of micro grain-boundary water-channels through which melt-water can permeate into the glacier body.

When a small amount of diluted ink was sprayed on each glacier's surface, it permeated down mostly along the planes of the grain boundaries or through the water-channels along the triple grain boundaries in the glacier body. The flow of the ink was clearly observed on the wall of a pit in the Mendenhall Glacier terminal zone.

A network of such water-channels was observed in a large ice block cut from the glacier. Melt-water contained in the water-channels was percolating out of the ice block from channels averaging 1 to 1.5 mm thick. Thus in this study the role of water channels is clarified with respect to the deep permeation of melt-water into Temperate glaciers. If such melt-water reaches the bed, it can play an important role in glacier flow and, via basal lubrication, even relate to the character and cause of glacier surges.

I. STREAM GAGE RECORDS FROM THE OUTFLOW OF THE CATHEDRAL GLACIER, 1972-75

To correlate with the 1972-75 hydrometric records on Lemon and Ptarmigan Creeks in the coastal valley of the Lemon Glacier System (Fig. 23), similar hydrometric data have been obtained in these same years from Cathedral (Elixer) Creek, the outflow stream of the Cathedral Glacier in the Camp 29 sector of the Atlin region. The locations of these two key research sites are indicated in the inset boxes in Figure 2.

The records discussed in Sections B, C and D above relate to the summer fluctuations in Ptarmigan Creek (as typified in Figure 30). The Ptarmigan and Lemon Creek discharge data may be compared with the records for the last half of summer from Cathedral Creek, as noted in Figure 48.

Of significance is that the Cathedral Glacier runoff has been abruptly terminated by freezing conditions in mid-September each year. For example, in 1972 this was accomplished when a blizzard brought the first winter snowfall to the Camp 29 sector on September 16. In contrast, in that autumn, the flow in Ptarmigan Creek continued well into October. Such is a reflection of the maritimity vs continentality affecting the duration of annual runoff from these two outflow streams on opposite sides of the range (as noted earlier in this report).

The field records on the Cathedral Glacier only cover the June to September period in each of the past four years, but measurements will be continued through this decade. Details of the study to date are given in several sections of this report as they are so closely allied to other basic facets of this regional terrain analysis.

J. THE EVOLUTION OF SUPRAGLACIAL LAKES AND STREAMS AND FLUTED ICE SURFACES ON THE GILKEY AND VAUGHAN LEWIS GLACIERS *

During the summer of 1970 to 1974 a number of observations and measurements were made on the Gilkey and Vaughan Lewis Glaciers with respect to ice surface changes related to the development of supraglacial streams. This research is still in progress and is allied with the survey and photogrammetry work briefly discussed in Part V. It extends studies on hydro-thermal erosion of glacier surfaces initiated by Dr. A.C. Pinchak in 1968-71 (Pinchak, 1972).

Supraglacial streams develop during the warm summer months as a result of melt-water produced by ablation. The melt-water which is not absorbed or vaporized then runs to lower elevations and in so doing develops tributary and trunk stream systems. In many cases, as on the Vaughan Lewis Glacier, deep canyons are cut in the ice with nearly vertical sidewalls. These walls have flutings or grooves running parallel to the stream bed (v. Fig.52). An investigation of the origin of these streams and the genesis of their fluted walls is briefly discussed below.

Research Locale and Investigational Methods

A well-developed multiple moraine system occurs at the junction of the Gilkey and Vaughan Lewis Glaciers noted in Figures 53 and 55a,b. The moraine is riven with a supraglacial stream system, highlighted in the figures by spicular reflections from the setting sun. On the left of the moraine is the Vaughan Lewis Glacier; to the right is the Gilkey Glacier. The large arrow (Fig. 55) shows direction of glacier flow, while the small arrow indicates the positions of the down-glacier moulins into which the streams plunge and disappear beneath the ice. The areas closely studied are shown in the circled sectors.

The drainage channels do not follow strictly the crevasse pattern. Many crevasses are water-filled, with the water level (perched water tables ?) 20 to 30 feet below the stream levels. The streams sometimes follow a crevasse which is fortuitously located as along a residual depression formed by the closure of a crevasse. Quite often the stream direction is independent of the crevasse pattern, be it present or relict. During the period 1968-1971, the stream systems were quite similar each summer. In 1972, stresses in the ice caused severe changes in the wave-bulge terrain at the base of the Vaughan Lewis Icefall, with the result that by 1973 and 1974 a whole new drainage configuration developed.

Visual readings were taken of water levels at recording stakes in three stream channels and a fresh pool. Ambient air temperatures were taken by rapid whirling of a conventional mercury-in-glass thermometer at the same time as the stake reading was recorded. In mid-summer, the water level lowered at about 4 cm per day.

Estimates of stream velocity were obtained by measuring the time required by a surface float to move a measured distance. Cross-sectional profiles were

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* This section is briefed from a manuscript report by Dr. A.C. Pinchak, Associate Professor of Fluid Mechanics, Case Western University.

obtained by steel tape traverses. Detailed measurements of grooves were taken with a vertical plumb bob and tape. Development of the grooves (with time) was directly observed by periodically measuring the vertical distance between the stream surface and reference pitons placed in the canyon wall.

Of the two main streams studied, one drained through the moraine in the zone between the Gilkey and Vaughan Lewis Glacier; the other from a portion of the area just below the Vaughan Lewis Icefall. The latter each year has connected several lakes impounded annually between the wave-bulges (Fig.53). The stream draining from the wave-bulge or ogive sector is referred to as the "ogive stream"; the other as the "moraine stream".

A levelling rod and theodolite were used to determine the longitudinal profile of the moraine stream. Reference stakes placed in each stream allowed the variations in water level to be observed and compared. Cross sectional profiles of the submerged bed of the moraine stream were delineated as well.

The Hydrological Analyses

Figure 56a and b shows typical variations in lake level, with air temperature and time of day. In order to relate a variation in pool level with stream flow, the storage effect of the pool had to be considered. The lag between pool depth and volumetric inflow rate was on the order of 5 minutes. Thus for our present purposes, the pool level is related to instantaneous flow.

Air Temperature and Stream Flow Lag

Figure 56a and 56b also reveals a lag between air temperature variations and stream flow. This effect is due to the time required for melt-water to seep along the surface until it enters the tributary stream system. The curve in Figure 56 results from calculation of the cross correlation function between the two time series: air temperature and lake level. The cross correlation is defined by the following expression (see Kisiel, 1969):

$$C(\tau) = \frac{1}{\Delta T}\int_{t}^{t+\Delta T}\left[L(t+\tau)-\bar{L}(t+\tau)\right]\cdot\left[\theta(t)-\bar{\theta}(t)\right]dt$$

where $L(t)$ is the time dependent lake level, $\theta(t)$ is the temperature variation, and τ is the time shift. In order that a cross correlation calculation produce a useful result, it is necessary that the correlated functions have a mean value of zero and that any long term trends or "drift" be removed. Data from 19 July through 21 July were utilized in the calculation of Figure 57 . For this period, the temperature displayed a steady trend so a constant, average value of $\bar{\theta}$ was calculated, $\bar{\theta}$ = 40.45°F. However, the lake level ($L(t)$) was fluctuating about a decreasing mean value. This trend was removed by a suitable function, $\bar{L}(t)$ having a slope of -0.057 in./hr. The cross correlation integration was carried out numerically on a UNIVAC 1108 digital computer.

Inspection of Figure 57 shows that the maxima and minima of C (τ) are well defined. The maximum correlation occurs with a time lag of 5 hours. For

the case of two, sinusoidal, diurnal functions, a 12-hour time difference between the maxima and minima in C(τ) is anticipated. The maximum occurs when the time shift (τ)results in the two signals being "in phase", whereas the minimum C (τ) occurs when the functions are "180° out of phase". Figure 57 indicates a time difference of 11.5 hours between the maximum and minimum values of C(τ). This last result provides additional support for acceptance of the aforementioned 5-hour time lag for the drainage basin observed here.

Helmers (1967) has recorded the stage variations in Thomas Lake, which lies below the Thomas Glacier on the Juneau Icefield. His data show diurnal fluctuations which are nearly sinusoidal in shape, but with the rising stage somewhat steeper than the falling stage. Maximum stage occurred at about 2100 to 2200 hours each day. As the peak air temperature usually does not occur after 1600-1700 hours in this area, a time lag of 4 to 6 hours is also suggested for the Thomas Lake data. Similarly, as noted elsewhere, a several hour time lag is indicated in the summer stage levels of the pro-glacial lake of the Cathedral Glacier near Camp 29.

Adams (1961) also measured the incident shortwave radiation on White Glacier in the Canadian Arctic and noted a correlation between this radiation component and the stream flow. His data indicate a somewhat shorter time lag (two to four hours) between the peaks in the shortwave radiation and peaks in the stream flow variation. Inspection of Table X indicates that, especially on clear bright days, shortwave radiation is the dominant heat transfer mechanism to the stream. And thus it would be expected that stream flow should correlate better with radiation than with air temperature, an observation borne out by our recent research on the glacio-hydrology of the Ptarmigan Glacier.* Unfortunately, a radiometer was not available during the course of this particular research on the Vaughan Lewis-Gilkey Glacier System.

Effect of Radiation Cooling

One of the more interesting aspects of the pool level variations observed here is the difference between the decrease of stage level during warm and cold nights (see Fig. 56). On warm, cloudy nights, the level would drop steadily with time until the meltwater, formed the following morning, began to enter the stream system. On clear nights, radiation cooling dropped the surface water temperature below freezing, and a layer of ice formed on the surface of the pools and the slower moving sections of the streams. The reference pool water level would then remain elevated until break-up of the surface ice the following morning. Coincident with the ice break-up, the pool level would then drop very rapidly (Fig. 56b). There was no evidence of frazil ice formation at any time during the observations.

Channel Downcutting

Research shows the mean drop in pool level which was caused primarily by downcutting of the outlet streams. Spot checks indicated that the lowering of the surface of the test stream or the main stream bore a one-to-one correspondence with the drop in level at the measuring stake. Thus, this shows a mean downcutting of about 1.5 inches/day (3.8 cm/day) for those streams which did not change their mean depth appreciably during the observation period. A change in the rate of downcutting which was observed on 15 July is a phenomenon still unexplained. However, this may be related to the loss of snow cover on this region of the glacier which is noted usually to occur about mid-July.

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* Discussed in detail in Part III, Section B and C.

TABLE X

CALCULATED COMPONENTS OF THE HEAT TRANSFER TO SUPRAGLACIAL STREAM SURFACES

(For streams on the Vaughan Lewis and Gilkey Glaciers under summer conditions)

T_w (°F)	T_a (°F)	V_{50} (ft/sec)	Rel. Humidity (%)	Cloud Cover (%)	$\dot{q}_r$ (Btu/$_{hr}$ft^2)	$\dot{q}_c$ (Btu/$_{hr}$ft^2)	$\dot{q}_b$ (Btu/$_{hr}$ft^2)	$\dot{q}_e$ (Btu/$_{hr}$ft^2)
32	40	5	75	0	80	7.3	-37.5	0
32	40	10	75	0	80	14.7	-37.5	0
32	50	5	75	0	80	16.5	-33.0	5.3
32	50	10	75	0	80	33.0	-33.0	10.5
32	40	5	75	100	20	7.3	-12.0	0
32	40	10	75	100	20	14.7	-12.0	0
32	50	5	75	100	20	16.5	- 4.0	5.3
32	50	10	75	100	20	33.0	- 4.0	10.5

Positive $\dot{q}$ values indicate heat transfer <u>to</u> the stream surface.

$\dot{q}_r$ = short wave radiation heat flux to stream

$\dot{q}_c$ = convective heat transfer

$\dot{q}_b$ = long wave, back radiation from stream

$\dot{q}_e$ = heat flux due to evaporation or condensation at stream surface

T_w = water surface temperature

T_a = air temperature at a height of 50 ft.

V_{50} = average wind velocity at a height of 50 ft.

The question now arises as to the mechanism responsible for this amount of stream bed ablation. Some relevant, approximate calculations have been presented in other publications (Pinchak, 1972). These estimates show that if the slope of the stream bed along the channel is sufficiently large, as near the lip of a moulin, then viscous dissipation may provide for a significant portion of the latent heat of fusion required to melt the ice on the stream bottom.

Field observations have consistently shown that the size of the parallel grooves in the canyon wall increase as the water flow accelerates as it falls over the lip of the moulin. This change in groove size indicates an enhanced rate of downcutting which is supported by an increase in viscous dissipation and an increased heat transfer which result from acceleration of the turbulent stream as it flows over the moulin lip. Over most of the stream course, viscous dissipation is not a significant factor.

Another approximate calculation (Pinchak, 1972) indicates that the water would have to be warmed a mere 0.005°C in the measuring pool to account for the observed rate of stream downcutting. This very slight degree of super-heat indicates why the water temperature of the pool was measured as 32°F (0°C) on the mercury-in-glass thermometers.

Grooves in the Canyon Walls

In addition to downcutting their channels, the glac[illegible] streams also scoured an interesting profile on the canyon walls. Of these so-called grooves or flutes, a detailed measurement from the canyon wall of the combined stream is shown in Figure 59. The opposite wall of the canyon was essent[illegible]ly a mirror image of this wall. Because the canyon was about 2 meters deep and covered by a snow bridge, the canoyn walls were not appreciably ablated after the grooves were formed. At the average rate of downcutting (3.4 cm/day), we see that Figure 59 represents almost a four-week record of downcutting.

By working upward from the stream level, it is possible to determine the former pool depth when a particular groove or cusp was formed. For example, the pronounced cusp, centered about a height of 10 cm, indicates that the [illegible]an water level in the pool corresponded to the day of 15 July. Inspection o[illegible] Figure 56a shows a definite increase in the runoff on that clear, hot day. In an analogous fashion, the lowest groove in Figure 59 correlated with th[illegible] increased runoff of 17 July. In addition, the correlation of increased runoff with flute formation was observed directly by repeated inspection of the stream channel on days of large runoff and the following days. In all in[illegible]ances, a new flute formed whenever the stream flow was appreciably increased for several hours and then returned to near the original flow-rate. T[illegible]s process of groove formation has been observed many times during all seven [illegible]eld seasons and has been corroborated by several visiting scientists who observed the process without previous bias. The question then arises as to the mechanism responsible for the formation of a flute.

If a constant stream flow condition were provided, a steady downcutting of the channel would result, but with smooth, vertical canyon walls. This canyon wall profile will be termed the "equilibrium cross profile". It is assumed that the water temperature is slightly above the ice temprature and has

also reached a steady value. Such a condition occurs when the water temperature does not change along the stream course and the net heat transfer to the stream surface must be utilized solely to produce phase changes at the air-water and water-ice interfaces. In this concept, a certain unique stream width and cross-section will be determined for each combination of total flow rate and bed slope. (Effects of ice density and embedded foreign material will, of course, affect the equilibrium profile.) A brief discussion of the "equilibrium cross profile" is presented below.

If the stream flow were suddenly increased, then the stream cross-section would begin to change so as to provide a larger width. As this sidecutting occurs, downcutting also helps lower the water level. The result would be a profile as shown in Figure 60a. Conversely, with a decrease in total channel flow, a narrower equilibrium profile will result (Fig. 60b). By combining a flow increase followed by a flow decrease to the original flow rate, a flute is formed, as shown in Figure 60c. Thus, the diurnal variations in flow rate are directly responsible for the formation of these flutes. With larger peak flows, deeper grooves are produced because of the more severe sidecutting. If a very large increase in a stream flow should result on any given day, the elevated water level modifies the shapes of previously formed grooves and may even obliterate those grooves which are inundated.

Inspection of the ice forming the canyon walls indicated the presence of foliation and structural patterns which were similar to those observed in other streams and crevasses in the local area. Because of these structural similarities which exist in the ice throughout the drainage basin, it is unlikely that the ice structure is a crucial factor in the determination of the shape of the canyon walls.

For the turbulent streams observed here the cross-section does not appear to be the optimum semi-circular channel which has the largest hydraulic radius of any open channel cross-section. Instead the width is relatively widened. It is interesting to note that the stream bed cross-profiles of these supraglacial streams are very similar to the profiles obtained in laboratory tests with streams eroding through non-cohesive sands (Leopold, et al., 1964).

Time Lag Between Moraine and Ogive Stream Flows

In contrast to the 1968 observations (Pinchak, 1972) which were concerned with the cross-correlation between diurnal air temperature and stream level variations, in subsequent seasons the research has concentrated on the relative time lag between two observed streams. Figure 61 shows the surface level variations of the two streams with an abnormally high run-off on 24 August in both streams. A comparison of these two peaks shows a lag of approximately 5 and 1/2 hours between the moraine and ogive streams.

In the cross-correlation calculation between the two stream levels (completely analogous to the air temperature-stream flow calculation), the data from the unusual day of 24 August were not included. A plot of this truncated cross-correlation which is in reasonable agreement is shown in Figure 58. Here a time lag of 5 hours is obtained which is in reasonable agreement with the single 5 and 1/2 hour estimate given above. It should be noted in Figure 58 that the maxima and minima of the cross-correlation repeat at average time intervals of 22.7 and 22.3 hours respectively. This result is in accord with the inherent diurnal variation expected in stream fluctuations.

It is reasonable to expect that the ogive stream flow should lag behind the moraine stream. This lag is produced primarily by the integrating or "smoothing" effect of impounded pools present in the ogive stream system. Such ponds are noted in the low points between wave bulges seen in Figure 53.

Conclusions

Supraglacial streams erode their channels by ablating the ice of their stream beds. Over most of the stream length only a small portion of the energy required to melt the ice is produced by viscous dissipation in the flooding stream; with the major fraction supplied by radiation and atmospheric convection. The degree of superheating (above 0°C) of the stream water is extremely small even on very hot, bright days. However, very small amounts of water superheat can account for the large amounts of stream bed ablation noted here. Heat transfer to the water is greatly increased as a stream passes over a moraine as evidenced by the enhanced rate of downcutting in and downstream of the moraine (Fig. 55a).

Diurnal fluctuations in stream flow rate and stream temperature cause a modification of canyon shape and thus produce a longitudinal groove pattern on the canyon walls. A new flute or groove is formed during the peak flow periods of hot, sunny days which have a relatively large runoff.

Variation in pool water levels was markedly different on clear nights as compared to cloudy nights. On overcast nights a steady decrease in pool level was noted. On clear nights the pool maintained a relatively high level until it rapidly dropped as the surface ice formation on the streams and pools was disrupted by the effects of the daytime warming.

A time delay between diurnal air temperature variation and stream flow rates was found in all streams. This time lag varied with the size of the drainage basin and the presence and size of upstream lakes or pools. Thus the flow rates of any two supraglacial streams will, in general, show a time lag of one stream with respect to the other.

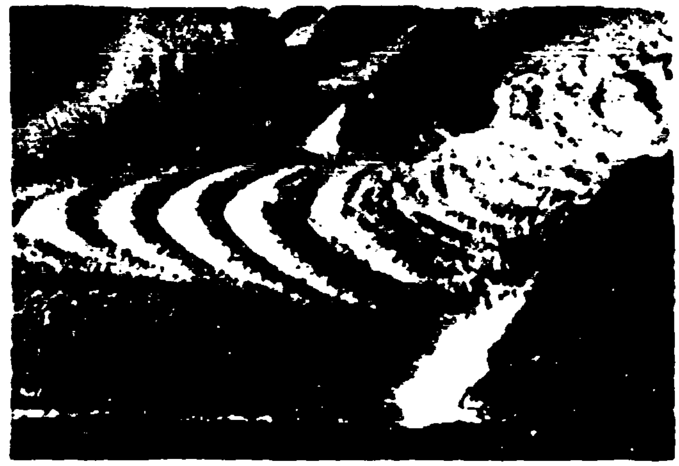
Longitudinal profiles of supraglacial streams ending in a moulin are, in general, convex upward. However, when such a stream passes over a moraine the increased heat transfer to the stream can produce a concave upward profile for a limited region through and just downstream of the moraine.

Pages 90, 91
a,b,c,and d

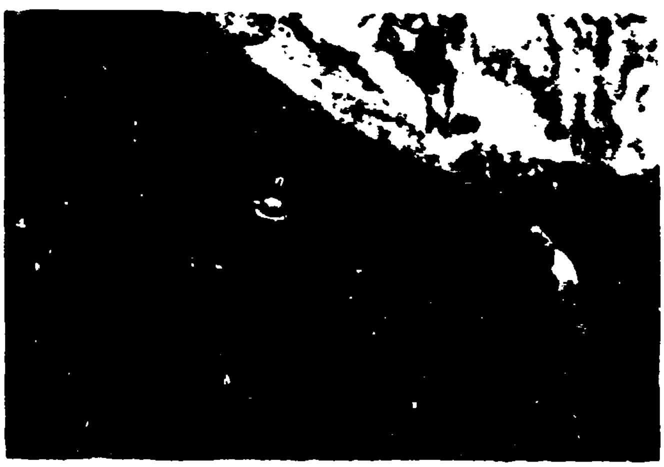
K. PERTINENT ILLUSTRATIONS

SELECTED VIEWS OF RESEARCH SITES
INVOLVED IN JIRP GLACIO-HYDROLOGY, GLACIOLOGY
AND PERIGLACIAL INVESTIGATIONS, 1971-74

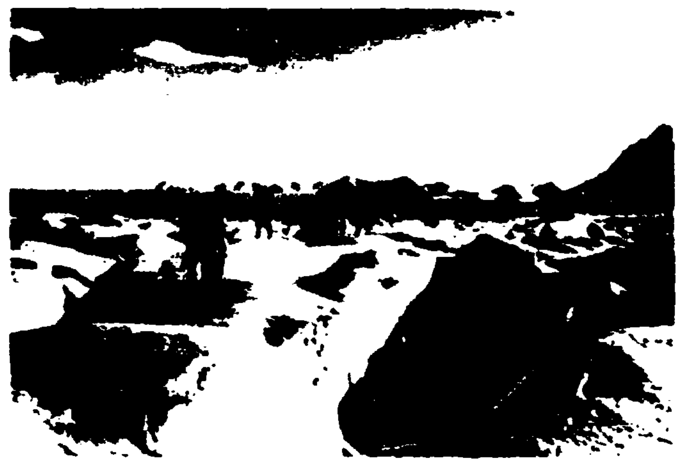
Moat-impounded supra-glacial lake in Icy Basin, Camp 10 sector, Juneau Icefield. (F.G.E.R. Photo, early July, 1972)



Upper Vaughan Lewis Glacier Icefall, with sections of Gilkey Glacier (above) and Whatchamacallit Glacier (below) (F.G.E.R. Photo, Sept. 24, 1972)



Debris-entrained basal ice at terminus of Mendenhall Glacier showing foliation structures and pitted zones from which entrained boulders ablated (early July, 1973)



On traverse of Llewellyn Glacier between Camps 26 and 27.

Developing palsa (frost mound) in esker complex at 3000 feet elevation in middle valley of Fourth of July Creek, Atlin region (Sept. 1973) Note toe of esker in immediate background.

Erratic boulder on late-Glacial moraine at 5000 feet near Camp 29 and Cathedral Glacier, Atlin region.

Small tor at 3500 feet elevation on Ptarmigan Ridge in Camp 17 sector (July 1973)

Relict stone circles on surface just outside of Neoglacial moraine sequence near Camp 17A. Center areas of circles thickly vegetated with heath matte. (July, 1973)

Tank topography at 3300 feet elevation on Ptarmigan Ridge (July 1973)

Ridge-top flat above Camp 17A, with relict stone circles along crest. (photo early July 1973)

Recent lateral moraine (1920's) on east wall of Ptarmigan Glacier Valley, as viewed in early July, 1973.

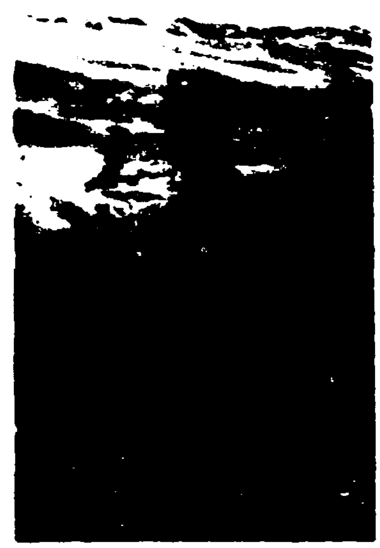
Lake Linda, a moat-dammed water-body on upper Lemon Glacier in mid-July, 1973. During spring and summer this self-dumping lake is a source of anomalous hydrological fluctuations in Lemon Creek.

Zone of flow foliation and tension crevasses below the neve-line on lower Lemon Glacier where some of the research on seismic anisotropism was conducted in 1973. (F.G.E.R. photo)

Tectonic (flow) foliation on Llewellyn Glacier with drag folding induced by cross-cutting fault (at top of photo).

Waves bulges on ice apron at base of Vaughan Lewis Icefall, near Camp 18. (Sept. 24, 1972, F.G.E.R. photo)

Lineated marble and lime silica granulate within migmatite zone of bedrock on Taku B headwall near Camp 10.

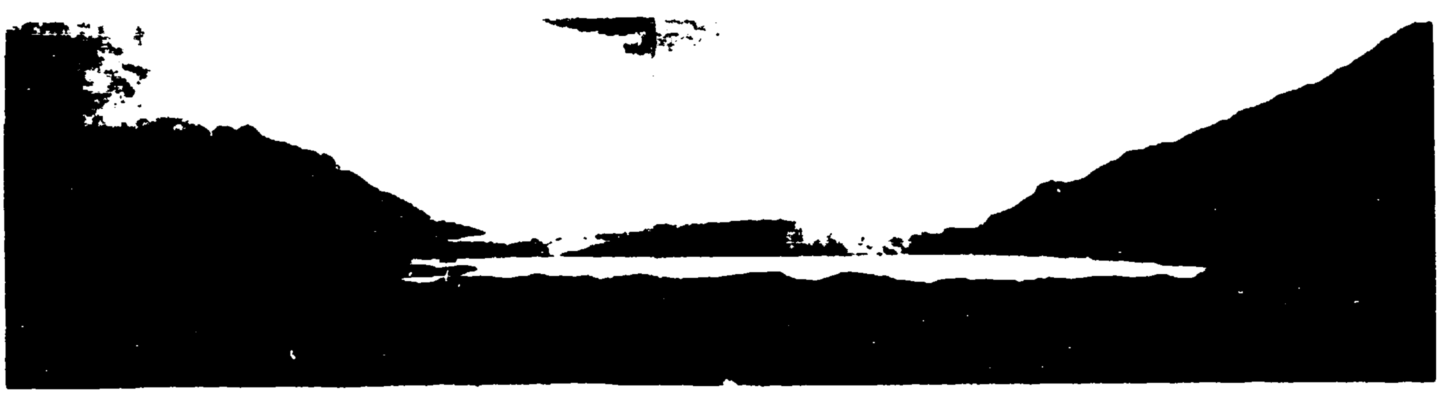


Inactive moulin on lower Llewellyn Glacier near 3800 level in late August, 1973.

91a'

Mendenhall Glacier terminus, Juneau Icefield, Alaska on July 9, 1973 (photo by M. M. Miller)

Advancing Taku Glacier terminus on July 7, 1973 (photo by M. M. Miller)



Receding Llewellyn Glacier terminus with pro-glacial lake and segment of outwash zone (valley train) August 21, 1973 (photo by M. M. Miller)

PART IV GLACIOLOGY

A. MASS BALANCE STUDIES

Detailed mass balance measurements have been conducted on the following glaciers covering the period 1971-75. Their current regime status is noted as well.*

	Semi-permanent névé-line(ft.)	Mass Balance
Mendenhall Glacier*	3400	slightly negative, trending toward equilibrium
Ptarmigan Glacier	3700	equilibrium
Lemon Glacier	3600	slightly negative
Taku Glacier*	2900	strongly positive
Hole-in-Wall Glacier	2900	strongly positive
Norris Glacier	3000	negative
Vaughan Lewis Glacier	4200	equilibrium
Llewellyn Glacier	4900	negative
Cathedral Glacier *	5700	slightly negative

Test pit data for determination of yearly net accumulation segments have been obtained on the Ptarmigan, Lemon, Taku, Llewellyn and Cathedral Glaciers over the same period. Névé-line information has also been obtained with the mean over this period as noted in the table above. The trend of the névé-line over the last 30 years on the Taku Glacier has been presented in Table IX.

In the firn stratigraphy, free water content measurements have also been made. As the mass balance records are so comprehensive, involving hundreds of crevasse-wall and test pit profiles over the past decade, only six typical profiles for 1972 and 1973 are given in Figures 50,62; Appendix G. Comparisons and analyses are left for other publication.

B. GLACIOTHERMAL RESEARCH

During the period 1971-74 thermistor cables were installed on the high crestal plateau of the Juneau Icefield (i.e., at 6000 to 7000 feet elevation near Mt. Nesselrode) as well as at the 6500-foot level on the Cathedral Glacier. Comparative measurements have been made to determine changes in englacial temperature over the past 20 years (see Miller, 1954b and Andress 1962). This research is continuing. Details will be published after two more years of comparative readings. It is clear from our studies, however, that both of these glacial areas are thermophysically sub-Polar to sub-Temperate (Miller, 1963, 1975c).

C. STATISTICAL STUDY OF ICE AVALANCHES

As reported in Pinchak and Lokey (1972) and Pinchak (1968), seasonal meteorological factors affecting avalanche frequency on icefalls of the Juneau Icefield have been under study. Serac avalanches were observed for varying lengths of time at the Vaughan Lewis Icefall. Frequent intermittent measurements of meteorological parameters, along with 24-hour a day observations by watch personnel, were subjected to rigorous statistical interpretation. The results yield a number of interesting conclusions with regard to the phenomenon of ice avalanches from the zerac zones of hanging glaciers, including changes in

* See photographic illustrations on preceding pages

their frequency and the effects of memory time in the parent ice mass. Differences in avalanche activity were noted with respect to clear and overcast skies and between similar sky conditions during day and night hours. The relative frequency distribution of avalanche magnitude demonstrates a smooth inverse relationship.

Calculated waiting times between successive avalanches were found to be exponentially distributed throughout each field season (1968-74). In addition, the same icefall memory time has been consistently observed in each summer. The non-Poisson distribution of avalanche frequency suggests a change in the Poisson frequency under the influence of changing weather, time of day and so forth. A particularly strong correlation is found with conditions of overcast sky and precipitation. The general tendency to increased avalanche activity during intervals of changing weather patterns as contrasted with the frequency of ice falls during periods of stable weather can be a matter of concern and safety to glacier travelers. Scientifically it is of interest because ice avalanches constitute an important element in the ablation and transportation of steep and highly crevassed glaciers.

D. AUTOMATED MONITORING OF AVALANCHE INTENSITIES*

As a corrollary to the avalanche frequency studies, in 1971 and 1972 more precise avalanche monitoring on the Vaughan Lewis Icefall (p. 91a) was carried out using automatic equipment. For this purpose, a light-weight electrically operated electronic device was developed for use in conjunction with a tape recorder to monitor background glacier noise. This included both ice flow noise and the flow of supraglacial water (Neave and Savage, 1970).

The field unit was a hydrophone of the type commonly used in sono-buoys with a sensing element having a sensitivity of -87 db referenced to a 1 volt/microbar. This hydrophone was lowered into water-filled crevasses or supra-glacial pools with care taken to avoid hydrophone-ice contact. Hydrophone depth was set in the range from 2-10 feet (0.6-3 meters). The associated electronic amplifier and tape recorder were located on an insulated pad on the edge of the crevasse or pool. The hydrophone signals were monitored on the tape recorder input at the beginning and end of each tape. Due to the sensitivity of the system it was necessary for the operator to remain very still during the recording period or to move to a position distant from the hydrophone. Serac avalanches were also observed visually and acoustically by the operator and recorded in a field notebook to facilitate review of the tape recording. Tapes were replayed in the field and later monitored visually on a cathode-ray-oscilloscope and a light-beam oscillograph at Case Western Reserve University.

An integral, potted preamplifier was connected close to the sensing element (7" or 18 cm) and had a gain of 75 db. Frequency response of the hydrophone and preamp increased from 5 Ha at 6 db/octave up to 15 KHz (personal communication, F. Hess, Woods Hole Oceanographic Institution). From this amplifier the signal was fed to a portable cassette tape recorder (Ampex Co., Model "Micro 70"). Overall response of the tape recorder varied from -db at 50Hz, +7 db at 125 Hz to -6 db at 7500 Hz. The entire system was powered by three 6 volt and one 9 volt batteries. Backpacking was facilitated as the complete system weighed less than 20 lbs. (9.1 kilograms) with batteries included. Total cost was less than $250. A qualitative idea of the system's sensitivity is its ability to detect a man walking on the glacier at a distance of approximately 200 feet (60 meters).

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* A paper on this research has been published in the Journal of Glaciology (Pinchak, 1972b).

Results and Discussion

When the hydrophone was located in the lower regions of the icefall, the signal intensities were markedly reduced as compared to signals received from crevasse pools in the ogive region. This difference was attributed to the poor transmission of waves through broken surface ice in the lower icefall as compared to the improved wave conduction through the compacted, solid ice which extends to the surface in the wave ogive region (L. R. Miller, 1970).

In the oscillator signals, by comparison with the background noise, avalanches were readily detected with a signal-to-noise ratio of approximately 20 db. In addition to the continuous background noise due to nearby supra-glacial streams, there were intermittent periods with repetitive pulsatile signals. Because of the high frequency content of these signals, it was concluded that they were due to a local disturbance in the region of the ogives. This type of signal was noted at several locations in the water-filled crevasses of the icefall-ogive region but was not heard in a glacial lake on another glacier on the Juneau Icefield. The source of this peculiar signal remains unknown but will be the object of future investigation. The signal may relate to microseismic activity as noted in Part VI.

In conclusion, this inexpensive, light-weight hydrophone-amplifier-tape recorder system is capable of detecting serac avalanches in an icefall. The system is also capable of recording noise related to glacier flow and deformation as well as hydrologic phenomena in the supra-glacial melt water. It will have considerable value in future research on icefall avalanches in glacier regions (Pinchak, 1972b). Applications of this technique include its use in areas where personnel movements may be hindered by avalanches.

F. MASS AND LIQUID REGIMEN OF THE CATHEDRAL GLACIER *

This section concerns an integrated study of the different phases of annual snow regimen in the Cathedral Glacier, from accumulation through ablation and run-off. Comparisons are drawn with snow depths at Camp 30 in Atlin and at Log Cabin, B.C., the nearest available locations where continuing records have been made. Ablation measurements on the Cathedral Glacier surface are also discussed, as are the elevations of the seasonal névé-lines since 1971. The nature of the glacier firn-pack is described, including a presentation of test-pit data, free water content (FWC) and diurnal patterns of drainage over the three-year period between 1971 and 1974. Finally, the results of this ablation and consequent drainage are considered, in terms of the total hydrologic run-off being confined to Cathedral Creek and to a far lesser degree to evaporation and infiltration of groundwater in azonal soils, moraines and vegetation matter near the ice.

Regional Snow-Pack Records

In Figure 63, snow depths are shown (in inches of w.e.), on April 1st at snow courses at Atlin (since 1964) and at Log Cabin (since 1960). These are monitored by the Water Investigations Branch of the Provincial Government of B.C. (Hydrology Division, 1971, plus periodic reports). Log Cabin is located at 2880 feet (880 m) elevation on the high plateau between White Pass and Carcross on the Yukon and White Pass Railway, at the north end of the Boundary Range (Fig. 3) Not only is Log Cabin at almost 500 feet (150 m) higher elevation

* Prepared with the cooperation of V.K. Jones, Dept. of Geology, Mich. St. Univ.

but it is in the path of maritime air masses which pass north-northeastward up Lynn Canal, over Skagway and on up the valley through White Pass.

Ablation Records and Nèvè-lines

Cumulative rates of ablation of the winter snow-pack on the Cathedral Glacier are given in Jones (1975,Fig.29The data are based on stake arrays, with measurements taken each evening. The measurement stakes melted out at 104 cm of total ablation on August 1, 1973. In 1974, the final stake in the lower group melted out at 129 cm of ablation by August 15. As expected, the rate of ablation was slower at the higher stake in 1974. The last 40 cm of firn at the lower sites melted at an accelerating rate, perhaps due to heavy infiltration of melt-water from above. The glacier surface at this point slopes steeply to the north, so that large quantities of melt-water from above saturated the firn-pack and later broke out as straight supra-glacial streams. Within a few weeks these streams had developed sine-generated forms indicating the rapidity at which the self-regulating glacio-fluvial processes strive for equilibrium on this smooth glacier surface.

These ablation records typify the negative mass balance statistics obtained on the glacier in each of the referenced years. This is further corroborated by the following record of relative average levels of the seasonal (late-summer position) and semi-permanent (mean lowest level over preceding 5 years) nèvè-lines, based on Camp 29's elevation of 5300 feet (1620 m).

Cathedral Glacier

	Seasonal Nèvè-line	Semi-permanent Nèvè-line
1971	5750 ft (1730 m)	5750 ft (1730 m)
1972	5800 ft (1760 m)	5700 ft (1735 m)
1973	5600 ft (1700 m)	5600 ft (1705 m)
1974	5700 ft (1735 m)	5700 ft (1735 m)
1975	5500 ft (1665 m)	5500 ft (1665 m)

The Local Firn-Pack

Englacial temperature measurements to a depth of 18 meters in 1972 indicated that this glacier is thermophysically sub-Polar in its upper head-wall portion (i.e. above 6000 feet, 1820 m). Below 5700 feet (1730 m) elevation, however, it is essentially Temperate (0°C). Thus surface melt-water freely percolates into the firn and flows off the glacier below the equilibrium line (because of the Temperate thermal character of the lower glacier, essentially the nèvè-line in this case).

Test-pit measurements at the same location as the FWC measurements (5850 ft., 1455 m) showed that the firn changed quite suddenly to bubbly ice at a depth of 130 centimeters. As the test-pit elevation was close to the mean elevation of the 1972 nèvè-line, the firn-pack depth is representative. This not only gives the annual net gain of the firn-pack (as Sept. 7 was close to the end of the ablation season) but suggests that in this glacier the firn changes to glacier ice at a relatively shallow depth. It further implies that this transformation may take place through re-freezing of melt-water into new ice (superimposed ice) above the previous glacier ice. It is significant that this glacier does not have a great deal of melt-water oozing out from its basal ice at the glacier toe (note Section III, H). These observations seem to corroborate the

sub-Polar thermophysical character of the headwall section of the glacier. It should be noted, however, that some surface drainage still takes place from this zone via selected channels along the glacier surface.

Liquid Water Content and Drainage

Free water content measurements (FWC) were made in 1972 using the calorimetric method in the Cathedral Glacier's surface firn pack (Appendix F). The FWC was measured at the following depths below the snow and firn surface.

Cathedral Glacier Firn-Pack at 5850 Feet (1455 m) on September 7, 1972

0 cm	Max. 13%
20 cm	5%
50 cm	5%
100 cm	Fluctuational

As shown in Figure 48, the FWC increased with solar radiation shortly before noon and reached the maximum peak noted above whether the firn was in shade or not. Also revealed is a strong relationship between FWC and ambient air temperature at the 5850-foot (1770 m) level. FWC decreased within one hour after solar insolation ceased. A decrease in the surface level of the pro-glacial lake occurred within 2 to 3 hours after reduction of liquid held in the firn-pack, showing good correlation between the changes in FWC and rise and fall of the lake. Thus changes in storage of liquid water and run-off in the Cathedral Glacier are shown to be dominantly controlled by variations in diurnal weather. The same may be expected with respect to longer-term climatic trends.

Hydrologic Run-off

Two types of hydrological measurements have been taken on the Cathedral Glacier over each summer, one since 1972 and the other since 1973. One, a water-level recording gage with a stilling chamber, was in operation during the 1973, 1974 and 1975 field seasons adjacent to a bedrock block on the northwest margin of the pro-glacial lake. While data from this gage cannot be analyzed in detail in this report, certain patterns emerge.

Firstly, substantial run-off does not begin on the Cathedral Glacier until early June, at which time the terminal lake is usually still frozen. Ice does not go off this lake until about June 20th as its elevation is relatively high, i.e. 5250 feet (1585 m). In fact, even at the relatively low elevation of Atlin Lake (elev. 2200 ft., 670 m) ice often remains in the coves and inlets until the first week in June.

At the terminus of Cathedral Glacier, a pro-glacial lake drains to the north. During high water, a vigorous outlet stream rushes through the rocky channel. When the lake is low, little surface flow occurs, but there is some sub-surface drainage through the moraine. Glacial ice from the upper cirque terminates at the up-valley edge of the lake (Figs. 8b and 9).

Higher run-off rates relate to the incidence of rainfall, increased solar insolation, and high ambient temperatures. On a good melting day with a

high p: tion of sunshine, the water level has been observed to increase sharply by noon (see Fig. 48). It peaks in mid-afternoon, and decreases in later afternoon even before the sun drops behind Splinter Peak. Such an early decrease in daily melting provides an interesting question. Does the drop in lake level during the night depend solely on temperature? If glacial surface temperatures drop below freezing, melting ceases, meltwater drains out of the firn-pack, and the lake level falls by early morning. On warmer nights, ablation and run-off continue and the lake level shows a less pronounced decrease during the night.

The other hydrological instrument employed in this study was a Stevens "A" stream level recorder installed with its float in a stilling chamber in the Cathedral Creek, some 300 years below the main Neoglacial terminal moraine (Fig.8b). Here again good correlation is found between ambient air temperature, solar radiation, ablation and run-off (Fig. 48), but during each of the past two summers this stream gage appears to have experienced peaks earlier in the day than the lake-shore recorder.

As the stream-gaging station is located about a mile down-stream from the lake-level recorder, this difference in time lag does not appear at first glance to be logical. The lake receives water from much of the upper cirque, and the stream flowing from this lake contributes a small but significant fraction of the total Cathedral Creek flow. A large proportion of the creek flow issues from an outlet at the glacier terminus, draining all the lower cirque and some melt from the upper cirque. A third but rather small component this late in the season comes from a few remaining lower-valley snow-patches and from slowly melting ice-cores still remaining in some of the moraines.

It is suggested that ablation near and above the seasonal névé-line in early September results in saturation of the firn-pack, producing a reservoir effect within the firn-pack. In the lower cirque, however, melt-water runs off the surfaces of the exposed glacier ice and drains away in well-established stream channels, taking short cuts through moulins and crevasses to basal streams. Thus it could drain out more quickly and reach the stream recorder down-valley sooner than water from the slushy firn-pack above the marginal lake.

The Cathedral Creek is an ideal closed system for hydrologic run-off and liquid balance studies. The entire run-off from the Cathedral Glacier system is restricted to the Cathedral Creek outflow. Therefore, any significant changes in ablation and precipitation are directly reflected in the rate of discharge into this creek. The seasonal discharge culminates in a final cessation of flow when the autumn freeze-up takes place. Over the period 1971-75, the freeze-up occurred in mid September.

Conclusions

Conclusions drawn from these observations are that in the present decade the effective ablation season on the Cathedral Glacier is of less than three months duration but usually greater than two and one-half months. In spite of the long winter season, the net mass balance of the Cathedral Glacier in 1971 through 1975 has been slightly negative. From the regional weather records this situation has apparently pertained since the 1920's. Thus it is imperative that future glacier mass balance and hydrometric records be acquired annually and without interruption in the years ahead in order to document future effects of the current cooling trend, which is expected to bottom out by the turn of the century.

Snow course accumulation data are as yet too short term to be conclusive, but the situation in 1974 suggests that patterns of snow depth on the Cathedral Glacier may be more closely related to Log Cabin than to the Atlin Lake snow course (Fig. 63).

Snow and firn-pack depths, ablation rates, late-season névé-lines, and the transient snow-line were all quite different in 1974 than in 1973 and 1972. In 1974 bedrock features and moraines were exposed to a much greater extent than in the previous two years, with much more pronounced down-wasting of ice in the terminal area.

Hydrologic studies are continuing on the Cathedral Glacier system. As this is a closed hydrologic basin, improved techniques and equipment will pay dividends in the mass balance studies and other aspects of the liquid and heat balance research.

It is obvious that interpretive problems arise from having too short a period of record at the Cathedral research station. Longer time-intervals are required to eliminate the effects of short-period variations and to reveal actual trends. But a start has been made in what is projected to be an important long-term program of glacial climatic and hydrologic research. As with Camp 17 on the Lemon Glacier, this station is a key in the synoptic studies of long-term regional glaciological and meteorological research. Hydrologically this area lies at the ultimate headwaters of the Yukon River. Therefore, these studies can play an important role in the hydrological balance and water management concerns in this region.

PART V PHOTOGRAMMETRY AND SURVEYING

MAPPING PROJECTS

During the period of this contract, maps were made of four key glaciers and one rock glacier , each of which are under comprehensive investigations. These are:

Lemon-Ptarmigan Glacier System: A composite large-scale (1:17,500) map was compiled of this system by computing information from several existing maps (USGS, B-1 Juneau Sheet, 1948; The Amer. Geog. Soc. Sp. Map, 1964). This map has been produced as Figure 9 in Miller, 1972b and has served as a basic reference for glaciological and hydrological research in the Camp 17 area.

Taku Glacier: A map of the Taku Glacier terminus, based on our phototheodolite surveys and at a scale of 1:15,000 was completed and published in Miller, 1974,p.206. Cartography was accomplished by Dr. L Knasovicky of the Dept. of Surveying Engineering, University of New Brunswick. *

Cathedral Glacier: Similarly, a P-30 phototheodolite map was made in 1972 by Dr. G. Konecny of the Cathedral Glacier near Camp 29. Further ground work was completed by Karsten Jacobsen, so that this map is now being rendered for publication at a scale of 1:10,000. It will be extremely valuable for the detailed research activities continuing on this inland glacier system (Fig. 9).

Vaughan Lewis Glacier System: A large-scale (1:5,000) plane table map was completed by Dr. C. Waag on the Vaughan Lewis Glacier and adjacent parts of the Gilkey and Watchamacallit Glaciers in the Camp 18/19 area. This map shows terrain differences in comparison with a similar map produced in 1966 (Kittredge, 1967). The latest map is reproduced in this report as Figure 67.

Atlin Mountain Rock Glacier: A phototheodolite survey of the Atlin Mountain rock glacier was begun in 1972, with further movement stake surveys completed in 1973 and 1974.

Annual movement stake surveys have also been completed in each summer on the Ptarmigan, Lemon, Taku, Vaughan Lewis, Llewellyn and Cathedral Glaciers. On the Taku Glacier, eight separate transects were re-surveyed for comparison with previous years (i.e., since such records began in 1949. Some of this work was accomplished in connection with gravity surveys on each of the glaciers noted (v. Section VII).

Each summer ground and aerial photographic documentation has also been obtained on the positions of termini of these and other key glaciers on the Juneau Icefield (e.g., Fig. 78 and photo supplement, pp. 90, 91).

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* Note: A July 7, 1973 view of the Taku Glacier is seen in the photo supplement in Appendix H.

PART VI GEOPHYSICAL STUDIES

A. GEOPHYSICAL SURVEYS OF SUB-GLACIAL TERRAIN CONFIGURATIONS *

During the 1971 to 1973 field seasons, gravity-depth surveys were carried out on the Lemon, Ptarmigan, Taku, Gilkey, Llewellyn and Cathedral Glaciers to determine the bedrock configuration beneath the ice, as well as glacier depths. In this report only the Lemon and Ptarmigan Glacier surveys are reported. The Cathedral Glacier study is being prepared for a subsequent report and the other surveys are part of a continuous geophysical project not yet completed.

Bedrock Geology of the Lemon-Ptarmigan Glacier System

The general geology of the area surrounding and underlying the Lemon and Ptarmigan Glaciers is best described as a transition zone between a sequence of progressively metamorphosed rocks (northwesterly trending and sub-isoclinally folded and overturned to the southwest) and the granodiorite to quartz-monzonite core area of the Alaska-Canada Coast Range Batholith.

Forbes (1959) has described the geology surrounding these two glaciers (Fig.23) as consisting primarily of alternating lime-silicate rocks and green schists, with some massive amphibolites and marbles on the western and headwall portions of the Ptarmigan Glacier. The remainder of the glaciers are underlain and bordered by migmatitic gneisses (quartz diorite and granodiorite plutons (Ford and Brew 1973).

For the purposes of the gravity survey, it is possible to group all of these lithologies together as country rock, with a mean density of approximately 2.86 gm/cc. This is due to the considerable overlap between density ranges of the main outcropping rock types.

Previous Geophysical Work

Previous geophysical work on the Lemon and Ptarmigan Glaciers has been quite limited, consisting of a single reflection seismic line and four gravity profiles all made on the Lemon Glacier. The four gravity profiles (Fig. 65) on the Lemon Glacier were conducted by the American Geographical Society in connection with JIRP as part of an IGY project (Thiel, LaChapelle and Behrendt, 1957). In the 1968 summer, Shaw (Prather et al, 1968) supplemented these surveys with a reflection seismic line (Fig. 65, line SA).

Field Procedures

The instrument used for the gravity measurements carried out in 1971 was a LaCoste-Romberg gravity meter. A gravity base station was established at the Camp 17 research station on the southern part of the ridge between the Lemon and Ptarmigan Glaciers. This base was tied to the U.S. Geological Survey's Alaskan gravity base network via helicopter loops to U.S.G.S. bases at the Juneau Airport and the Douglas heliport.

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* Prepared by Richard M. Shaw, Exploration Dept., Exxon Company, and William J. Hinze, Dept. of Geosciences, Purdue University

The stations for the survey were laid out in the form of profiles, seven on the Ptarmigan Glacier and two on the Lemon Glacier (Fig. 65). Station spacing along the profiles was normally 500 feet with some intermediate 250-foot stations. Horizontal and vertical control for the stations was obtained via Wild T-2E survey's from existing JIRP survey bases.

Reduction of Observations

Standard gravity data reduction techniques as outlined in geophysics texts (Dobrin, 1952; Parasnis, 1966) were used to reduce the observed data from the survey. All stations were corrected for time variations (drift), latitude and mass (density of 2.76 gm/cc.). The resulting simple Bouguer gravity anomaly data were than subjected to terrain corrections (Hammer, 1939) using a density of 2.76 gm/cc to remove the effects of the nearbv large peaks and valleys. The resulting data were then plotted in profile form for analysis.

Analysis of Results

The terrain-corrected Bouguer anomaly profiles were first analyzed by fitting simple two-dimensional gravity models to them (Talwani, 1959). The assumption of two-dimensionality for the profiles is, of course, only grossly valid because the causative bodies (the glaciers) are not infinitely long at right angles to the profiles. However, as a first approximation the two-dimensional technique works very well.

The gross configurations arrived at by the simple modeling were modified using end corrections (Nettleton, 1940) to account for the gross non two-dimensionality of the models. The resulting cross-sections are presented in Shaw et al, 1975.

The thickness information from the models has been combined with the previous gravity and seismi studies to prepare a revised isopach map of the Lemon Glacier and to construct an isopach map for the upper Ptarmigan Glacier from which the bedrock configuration can be determined (Fig. 66).

Conclusions

Analysis of these gravity surveys and sub-ice bedrock configurations of the Lemon and Ptarmigan Glaciers suggests that the névé surfaces do not closely reflect the true nature of the rock topography and reveal that a glacier in a separate cirque at the head of Lemon Glacier has merged with the main glacier. Similarly, three thick ice areas in the upper Ptarmigan Glacier indicate a complex merging of glaciers via multiple cirques to form the present Ptarmigan Glacier. Such information is critical in hydrological assessments, as in Part IIIa.

Additional work planned for the lower Ptarmigan Glacier and the upper Lemon Glacier when combined with this geophysical research will further clarify and amplify the results of these investigations.

B. SEISMIC ANISOTROPY BELOW THE ICEFALL OF THE VAUGHAN LEWIS GLACIER*

To test the degree and type of velocity anisotropy displayed by a Temperate glacier with well developed foliation, a seismic experiment was conducted on the Vaughan Lewis Glacier on the Juneau Icefield in association with other on-going research supported by the Army Research Office in the Camp 18 area (Fig. 1) In this field work a Geo-Space interval timer was employed, with four three-dimensional geophones arrayed on the surface of exposed bubbly glacier ice at some distance below the névé-line. The study zone was on relatively smooth glacier surfaces a short distance down-glacier from the lowest pronounced wave-bulge (wave-ogive) below the ice apron of the Vaughan Lewis Icefall (Fig. 53). In this zone the width of the Vaughan Lewis Glacier is nearly half as great as at the base of the icefall. The ogives (or arched bands) in this area are also more curved, as viewed from above, than further up-glacier, suggesting that the center of the glacier is flowing more rapidly than the edges at the medial moraines. The resulting shear stresses may be partially responsible for any observed velocity anisotropy as well as influencing the recrystallization of ice which would also effect the velocity field.

Petrofabric Relationships

Petrofabric analyses on other glaciers by Rigsby (1951, 1960) and Gow (1963, 1964) show one, two, three and four c-axis orientations on the Schmidt diagram in stressed ice. Steinemann (1958) has shown recrystallization during deformation to occur with the basal planes of ice crystals oriented in the direction of principal shear stress applied in the laboratory. Rigsby (1960) also shows that the c-axes in hexagonal ice crystals of a glacier tend to align from random orientation to a direction perpendicular to the shear stress after as little as two months of applied stress. Field studies of bore-hole samples at the 3600-foot (1090m) level in the Taku Glacier by Bader and Wasserburg have demonstrated that in successively deeper samples, below 140 feet, there is a progressive crowding of c-axes toward the normal to the presumed direction of main down-glacier flow (Miller, 1957, 1963, p. 132).

It has been shown in other glaciers that such fabric or foliation alternate with zones of bubbly and clear ice (Allen, et al., 1960; Shumsky, 1964; Paterson, 1969), with the planar direction nearly normal to the preferred c-axes orientation of Temperate glacial ice. Kamb (1959, 1961) has shown that the glide direction and direction of shear stress shall never be more than a few degrees apart. Paterson (1969) points out that the ice crystal deforms by gliding on its basal plane. Therefore if the stress is long-term and large enough to produce foliation, the alignment of c- axes in polycrystalline ice probably approaches directions normal to the foliation plane. This conclusion has also been discussed with respect to the Vaughan Lewis Glacier by L.R. Miller (1970). However, no successful petrofabric measurements have been made in the study area and thus we cannot be sure of the c-axes orientations. It is reasonable to expect, however, that the c-axes have a preferred orientation normal to the foliation and, in addition to crystal orientation, other factors may influence the observed velocity anisotropy such as structural effects and the presence of a stress field.

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* Report by Dr. Hugh F. Bennett and Barry W. Prather, Department of Geology, Michigan State University, research affiliates of the JIRP in 1972.

Field Proecdure and Experiments

The equipment used in this study included a Geospace GT2A interval timer with twelve traces and 10 millisecond timing lines recorded on Polaroid film. The time picking was accurate to 1/4 millisecond and repeatable to 3/4 milliseconds. The total record length was set at 0.15 seconds for best accuracy. The recording detectors were Hall-Sears 3-component geophones with a 4.5 Htz resonance frequency and with 62 per cent damping the system had a flat response from 7 to 125 Htz. The geophones were oriented with the X component toward the shot (horizontal longitudinal), the Y component perpendicular (horizontal transverse), and the Z component vertical. One pound charges of Nitramon Primer were used as the seismic source placed at 1 meter depths. A detailed explanation of the field procedures and some preliminary analyses have been presented in a fuller report by Prather (1972).

The seismic caps used gave a measured time standard deviation of $\pm$ 0.6 milliseconds. Combining this figure with the repeatability of 0.75 milliseconds gives an estimated time measurement accuracy of about $\pm$ 0.96 milliseconds. Since the time interval measured was approximately 50 milliseconds, the estimated accuracy of the P-wave measurements was less than 2 per cent. The same estimate is given for the shear wave measurements. The distance was measured to better than 0.1 meters with a steel tape and over a distance of 166 meters gave an error of less than 0.07 per cent. Even though the accuracy is not as good as we would like, since the P-wave velocity variation was only 4 per cent, when the data are fitted to the Q-ellipsoid, the results statistically indicate that the material measured is anisotropic. Furthermore the velocities were determined over a 4 geophone array and averaged so that the actual accepted velocities were probably good to $\pm$ 1 per cent.

Confirmation that the seismic ray paths travelled along straight, near-surface trajectories was obtained by a standard refraction profile oriented in the down-glacier direction. The results of this profile are shown in Figure 68 and show that the P-wave velocity (3670 m/sec.) is constant over the 100-meter distance. Therefore no velocity gradient is detectable and the ray paths sample the near-surface material as required in our experiment.

The ice thickness in the study area is about 200 meters so that no early arrivals from bedrock refractions are present and would arrive beyond the end of the record, if at all. For example, assuming an upper limit bedrock P-wave velocity of 6,000 m/sec. and ice P-wave velocity of 3,500 m/sec., the critical distance is about 775 meters. This, coupled with the refraction line data, indicates that we are sampling a single layer in the experiment over distances that are less than 200 meters.

The seismic array used in the anisotropy experiment is detailed in Figure 69 and shows that the velocities were measured over a 62° change in azimuth. The distance from shot point to the first detector ranged from 146.5 meters to 166.0 meters and the four geophones were on a line pointing down-glacier and spaced at 9.1 meter intervals. We were limited in our azimuth range by a medial moraine on the north edge of the array and by fracture patterns on the south. For each shot point the geophones were reoriented so that the x component always pointed toward the charge.

The study area, with respect to the entire glacial valley system, is shown in Figure 70a. The orientation of the Q-ellipsoid is also shown on this figure and will be discussed later. Note, as previously mentioned, that the foliation curvature is greater in the down-glacier direction and the width of the Vaughan Lewis Glacier is about half as great in the study area as compared with the width at the base of the icefall. This may indicate that the ice is undergoing compression from the sides since several glaciers are flowing contiguously down this narrow valley. Although the ice thickness profile along the axis of the Vaughan Lewis Glacier is unclear at present, it probably does not change very rapidly in the down-glacier direction from the base of the icefall to the study area (Fig. 70b)*.

Analysis of Seismic Data...Fit to Q-ellipsoid

The velocities of the compressional wave (P-wave), the vertically polarized shear wave (Sv) and the horizontally polarized shear wave (SH) were determined from the seismic array shots which were recorded on Polaroid film. A typical record (shot 25) is found in Figure 71 and shows the X, Y and Z traces for geophones 1 through 4. The P-wave is obviously the first arrival and is best recorded on the X and Z components as would be expected from theoretical considerations. The P-wave arrivals are easiest to pick and the transit time was measured from the shot instant to the point where each individual trace deviated from zero displacement (first motion).

The shear wave velocities were more difficult to determine precisely. In Figure71, the Rayleigh wave arrivals are indicated on the Z component traces and are also seen on the X component. The precursor just ahead of the Rayleigh wave is interpreted to be the vertically polarized shear wave (SV). By comparing the SV and Rayleigh wave arrivals on all records, except 21, a mean velocity ratio of $V_R/\bar{V}_{SV}$ = 0.933 was computed. This compares favorably with Knopoff (1952) for a Poisson's Ratio of 0.33, the calculated value from $\bar{V}_p$ and $\bar{V}_s$ in our study area. The calculation for Poisson's Ratio is based on the assumptions that the material is linearly elastic and isotropic. The SV wave velocities were thus computed by dividing the Rayleigh wave velocities by 0.933 for all seven records. This method was used because the Rayleigh wave arrivals were more uniform and easier to pick. Although this method may introduce a slight error to the absolute velocity calculation the effect is negligible on relative velocities as related to azimuth in this study.

The horizontally polarized shear wave (SH) arrivals were the most difficult to pick. Although it is difficult to prove theoretically that SH energy is excited by an explosive charge, it has been reported by many workers that SH energy is observed from nuclear devices on a large scale. Clearly, on these records, there are large amplitudes on the Y traces corresponding to the appropriate time for the Love Wave train arrival. It was assumed that these arrivals on the Y traces were the beginning of the Love Wave train and thus correspond to the SH arrival. Records 23, 24 and 25 are extremely poor for SH wave arrivals and the picks are not obvious. Because this study required only relative wave velocities, greater care was taken to pick the same event on each record rather than make certain that the first energy of the SH wave was picked. Furthermore, the SH wave data contribute only about 17 per cent to the value of Q so that errors in SH velocities do not seriously hamper the analysis. The accepted values for the P-wave,

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* Ed.: A reasonable assumption from an approximate sub-glacial profile with depths (800 ft.) drived from only a few seismic shots (Kittredge, 1967).

SV-wave and SH-wave velocities in this study are listed below.

Shot Number	Angle From Foliation	Velocity (meters/second) P-Wave	SV-Wave	SH-Wave
20	7°	3559	2000	1876
19	18°	3556	1923	1833
21	29°	3611	1943	1869
22	40°	359[illegible]	2037	1868
23	49°	3647	2050	1895
24	58°	3667	1965	1897
25	69°	3697	1948	1949

The velocity data from this listing were then tested with the Q-ellipsoid method described by Bennett (1972a). The Q-surface is a theoretical surface defined for single crystals by the formula:

$$Q = V_{P}2 + V_{SV}2 + V_{SH}2$$

where

V_P = P-wave velocity
V_{SV} = SV-wave velocity
V_{SH} = SH-wave velocity

The Q-ellipsoid is a triaxial representative ellipsoid whose value for any direction is $1/\sqrt{Q}$. Calculating the measured values of Q in the 7 directions for which the velocities were measured gives us

$$M_i = V_{P_i}^{\ 2} + V_{SV_i}^{\ 2} + V_{SH_i}^{\ 2} \quad (i = 1, \ldots, 7)$$

The values were determined over a planar surface so we thus determine the best fit ellipse to the measured values after the method of Nye (1957, pp. 163-168). The measured values are shown along with the best fit Q surface in Figure 72. The semi-major axis of the ellipse has a Q value of $21.7 \times 10^6\ m^2/sec^2$ and the semi-minor axis $19.9 \times 10^6\ m^2/sec^2$. The ellipse is oriented such that the minor axis is only 4° from the foliation.

The Q-ellipsoid in single crystals has an identical orientation with the optical indicatrix for hexagonal through monoclinic systems. Thus the orientation of the Q-ellipsoid indicates the principal directions to which the elastic properties of a material are referenced. If the material is uniformly anisotropic over the study area, it behaves as a single unit or as a pseudo single crystal and the Q-ellipsoid orientation gives us the orientation of this pseudo single cyrstal. In a sense this geophysical method allows us to do what is in effect a petro-fabric analysis over a large area in a rapid and simple fashion.

Criteria for Homogeneous Anisotropy

Now let us discuss the evaluation of the type of anisotropy found in the study area by the use of the standard deviation used in statistics. The three types

of deviations used in this evaluation are:

$$\sigma_{\bar{M}} = \left[\sum_{1}^{7} \frac{(M_i - \bar{M})^2}{7}\right]^{1/2} \qquad \text{where} \quad \bar{M} = \sum_{1}^{7} \frac{M_i}{7}$$

$$\sigma_{\bar{M}Q} = \left[\sum_{1}^{7} \frac{(Q_i - \bar{M})^2}{7}\right]^{1/2}$$

$$\sigma_{Q} = \left[\sum_{1}^{7} \frac{(Q_i - M_i)^2}{7}\right]^{1/2}$$

M_i = Value of M in the i th direction

Q_i = Value of Q in the i th direction

The first term, $\sigma_{\bar{M}}$, is the standard deviation of the measured values and can be thought of as the deviation of the data from the best fit circle in our study. The second term, $\sigma_{\bar{M}Q}$, is the deviation between the best fit circle and best fit Q-surface for the same suite of directions i. This value gives a measure of the amount of anisotropy. The third term, σ_Q, is the deviation of the data from the best fit Q-surface. If $\sigma_{\bar{M}}$ is greater than σ_Q the data fit the Q-surface better than a circle and the material is anisotropic. However, the anisotropy could be caused by inhomogeniety throughout the study area and not by a homogeneous or uniform anisotropy as displayed by single crystals. Anisotropy is defined in our study as a change in physical properties with direction and therefore could be caused by a directional change in homogeneity as well as uniform anisotropy throughout the study area.

If, however, $\sigma_{\bar{M}} > \sigma_{\bar{M}Q} > \sigma_Q$, the material is judged to be homogeneously anisotropic or centrosymmetric as is the case for single crystals. When this criterion is met we say that the material behaves as a pseudo single crystal. The criterion for pseudo single crystal behavior merely states that the degree of anisotropy, as indicated by $\sigma_{\bar{M}Q}$, must be greater than the scatter in the data from the Q-surface. This is a reasonable criterion and does not say that no inhomogeneity is present, but rather that the amount of homogeneous anisotropy is greater. In our study area the calculations are:

$$\sigma_{\bar{M}} = 5.13 \times 10^5$$

$$\sigma_{\bar{M}Q} = 4.68 \times 10^5$$

$$\sigma_Q = 2.11 \times 10^5$$

Since $\sigma_{\bar{M}}$ will always be greater than or equal to σ_Q, the value is an indicator of the degree of inhomogeneity in the sample, but also includes

measurement errors. As stated earlier σ_{MQ} is a measure of the degree of anisotropy. Thus the degree of inhomogeneity is probably best described by the value of $\sigma_Q - \sigma_{MQ}$. When this value is positive, heterogeneity predominates (or large measurement errors), but when the value is negative, homogeneity predominates. In our study, the value is negative and thus the degree of anisotropy is great enough to overcome errors in the velocity measurements as discussed earlier. The criteria for homogeneous anisotropy are met and the material in the study area is therefore judged to be pseudo-single crystalline in its mechanical behavior.

Conclusions and Summary

From this experiment we conclude that in the study area on the Vaughan Lewis Glacier approximated by an equalateral triangle, 160 meters on a side, the glacial ice behaves mechanically as a pseudo-single crystalline material to seismic waves. This homogeneous anisotropic unit may be primarily due to a strong alignment of ice crystal c-axes normal to the foliation which was produced at an earlier time, probably at the base of the icefall farther up-glacier. The P-wave, SV-wave and SH wave velocities do not fit the velocity models calculated by Bennett (1972b) based upon uniform distributions of ice crystal c-axes within a cone or on the surface of a cone with various apex angles. Therefore, it is possible that the c-axes may have a more complex distribution with a tendency to align normal to the foliation. On the other hand, the anisotropy may be due not only to the gross crystal orientation but be partly controlled by oriented structures in the ice and by the effects imposed by the presence of a stress field. Since petrofabric data are not available in this area at present, we cannot be sure how large a contribution each of these three factors makes. However, based on petrographic studies in other glacial areas, where foliation is pronounced, it is probable that strong c-axes orientations are the most important single factor.

Extending this analysis to other areas, we should expect, for example, that velocity anisotropy would be found along the Vaughan Lewis Glacier for some distance down-glacier where foliation bands are still detectable to the eye*. Where the foliation becomes nearly parallel with the medial moraines, however, the Q-surface major axes is probably orthogonal to the flow direction. There is no compelling reason to expect velocity isotropy to be the rule, especially in the flowing glaciers, and it is probable that most glaciers display varying degrees of velocity anisotropy between different glaciers as well as along the axis of any particular glacier. Because the field method used provides a reasonably fast and simple method of doing a petrographic type analysis over a large area with *in situ* measurements, we believe that it has great potential in the area of glacial geophysics as well as in analysis of other rock types.

The objective of this study was to test for velocity anisotropy on a highly foliated Temperate glacier, the Vaughan Lewis Glacier, located on the Juneau Icefield, Alaska. The seismic experiment included a standard on line refraction profile to confirm that the ray paths were direct near surface travel paths in the anisotropy measurements. The velocities were measured over a 62° change in azimuth using a straight line group of four 3-component geophones and a series of 7 shot points located along an arc. The compressional wave, horizontally polarized and vertically polarized shear waves were all found to have velocities which depended upon direction.

* v. Figure 70b

When the data are fitted to the theoretical Q-ellipsoid, the highly stressed glacial ice in this particular study behaves sonically as a pseudo-single crystal which has homogeneous anisotropy. The compressional wave velocity anisotropy is observed to be about 4 per cent with the maximum in a direction roughly normal to the foliation. The Q-ellipse, which represents the entire velocity data, has its semi-minor axis oriented approximately 4° from the foliation and thus its semi-major axis is nearly orthogonal to foliation. The main cause of the anisotropy is probably due to gross ice crystal orientation with a tendency of the c-axes to align normal to the foliation. However, additional factors which may influence the anisotropy are gross orientation of structural features in the ice and the presence of a stress field. It is apparent that the Q-ellipsoid method of analyzing directional seismic velocity data has many applications in glacier geophysics and in the study of other rock types.

C. ELECTRICAL RESISTIVITY AND GRAVITY SURVEYS

During the course of this contract, four projects were developed for the application of electrical resistivity methods in an attempt to delineate (1) water depths in pro-glacial valley train deposits of the Ptarmigan Glacier; (2) to differentiate firn-depths on the Lemon and Taku Glacier; (3) to explore ice-cored moraines in the Llewellyn and Cathedral Glaciers and (4) to determine ice depths in palsas and peat plateaux of the Atlin region. The latter study has been reported on in publication by Tallman (1973). Other aspects of this research are reported by Tallman in her Ph.D. thesis (1975a) and in the report of the Arctic and Mountain Environments Symposium held at Michigan State in 1972 (v. Tallman, 1975b).

More strictly glaciological research was carried out at an earlier date by Dr. Heinz Miller and during the period of this contract by Dr. Richard Kellogg whose reports are in preparation. Dr. Heinz Miller advises that on the Taku Glacier this method detected a unique interface at 40 meters depth in the firn-pack of the intermediate elevation (geophysically Temperate) zone in the vicinity of Camp 10. This presumably has significance in terms of the water content of the firn-ice material to be expected at this depth. We anticipate his final paper on this study.

Successful gravity depth surveys of the lower Gilkey and upper Cathedral Glaciers were conducted in 1972. The reports on these await completion of the photogrammetric map noted in Part V.

D. EXPERIMENTAL DETECTION OF GLACIER NOISE AND DISCONTINUOUS MICRO-MOVEMENTS

Geophysical techniques were applied to the detection and monitoring of glacier noise through the use of hydrophones and portable seismographs. Details of the hydrophone experiment are given in section IVD.(Also v. Pinchak, 1972b)

Seismographic Recording

In 1973 an experimental program was initiated using seismograph equipment to detect and record glacier "noise" A network of four portable seismograph units was emplaced in the wave bulge sector of the ice apron at the base of the Vaughan Lewis Glacier. Records were made on rotating smoked drums which, after a few days of operation, revealed a vast number of previously undetected microseisms attributed to minute increments of discontinuous shear-flow deep within the glacier. This interesting project was carried out with the help of Dr. Robert Page and William Gawthrop from the U.S. Geological Survey's National Center for Earthquake Prediction in Menlo Park. Their involvement was prompted because of a mid-summer earthquake registering 7.2 on the Richter scale, with its epicenter 50 miles off-shore from Cape Spencer, Alaska ... i.e., 175 miles west of Camp 18. In field studies of the after-shock, unexplained minor shocks were detected, believed possibly to relate to glacier noise in the La Perouse Glacier near the USGS field recording site. It was thus desirable to run some tests to identify the signature of glacier noise.

The microseisms measured on the Vaughan Lewis Glacier proved to be of particular interest to us. We believe them to relate to in-ice basal and marginal shearing, particularly in the compressive flow zone at the base of the icefall (Fig 53).In effect, the signatures corroborated our suspicion that all glacier flow is fundamentally sporadic and discontinuous.

Field mapping of folded basal foliation structures in this ice apron zone also proceeded during the 1972 to 1974 summers with their configurations found to relate to a pattern similar to that shown schematically in Figure 72. Earlier research on this subject has been reported by Miller (1968, 1975a). The continuing investigations on this problem during the period of this contract have been closely allied to the continuum mechanics and structural glaciology studies discussed in Part VII (following) and in the geophysical anisotropism research discussed in the preceding section (VIB).

PART VII CONTINUUM MECHANICS AND STRUCTURAL GLACIOLOGY*

A. Glaciers as Structural Models

Our various investigations on the Juneau Icefield have indicated that the Vaughan Lewis, Gilkey, Herbert, East and West Twin and other glaciers present an excellent opportunity for study of glaciers as structural models both geometric and kinematic (Waag, 1972).

Structures analogous to those found in plutonic rocks are abundant and include schlieren, schlieren domes, marginal fissures (Federklufte), cross-joints (Q joints or Querklufte) and stretching surfaces. In addition, along the margins where the glaciers are in contact with drift or relatively unconsolidated debris, marginal "upthrusts" originating as either Riedel shears or Federklufte, or both, can be identified.

The various glaciers within the Gilkey Glacier Trench also differ in their overall configuration of foliation. In some, the foliation is arranged in smooth curves convex down glacier and approximating a parabola (Fig. 70b). In others the foliation outlines complex subparallel digitations resembling shear folds or passive folds in other metamorphic rocks, and a wide variety of fold styles between the two end members exists (see photos p. 91a, c).

The causes and kinematic differences manifested in these differing styles are not yet understood, so detailed further studies are planned. Some detailed mapping, however, of these englacial structures has been accomplished. These efforts ally to the continuum mechanics investigations on these same glaciers and elsewhere on the Juneau Icefield. The results of these combined studies, when correlated with structural differences within the glaciers and their surface manifestations, will yeild useful models for understanding these structures, the morphological expression of them and the fundamental causal factors involved via the deformation stresses in glaciers as a whole. As examples of the specific kinds of studies involved, the following research is reported.

B. Firn Folds, A Model for Cover Rock Deformation

Investigation of deformation within firn and old snow overlying the ice of the Vaughan Lewis Glacier in the Gilkey Trench suggests a possible model for accomodation of cover rocks during basement shortening. Deformation occurs where the Gilkey Glacier complex swings through a right angle turn into the main trench and the icefall of the Vaughan Lewis Glacier contributes large quantities of ice (Figs. 53, 70a and b). A decrease of moraine width and convergence of the medial moraines through the turn adjacent to the base of the icefall indicate lateral shortening attendant to down-glacier movement.

Foliation and banding within the glacier ice strike parallel to the flow direction and generally dip steeply toward the icefall of the Vaughan Lewis Glacier. The decrease in width between the converging moraines is apparently compensated in the ice by an increase in depth and velocity. The change in shape is accomodated by flowage parallel to the foliation (Fig. 73), while initially horizontally bedded snow and firn accomodate by buckling (Fig. 54) into

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* Prepared with the cooperation of Dr. Charles Waag, Dept. of Geology, Georgia State University, W.A. Dittrich, Physics Department, University of Colorado, and Dr. Gordon Warner, General Motors Institute, Flint. The rotated foliation theory is after M. Miller (1975a).

concentric and disharmonic folds. Within a distance of 100 meters, the glacier decreases in width from 55 to 35 meters. This represents approximately 35 per cent shortening. The folds generated by the shortening range from 35 cm to 5 m in width, from 12 cm to 1.5 m in height and from 1.5 m to 25 m in length. The folds range from upright open forms to tightly appressed overturned forms, however, the direction of overturning is not consistent. The total area of upbuckling (anticlines) is small compared to the area of non-upbuckling (synclines). (Figs. 54 and 73b from Waag, 1974; also v. Pinchak, 1973).

C. Rhombus and Rhomboid Parallelogram Patterns on Glaciers as Natural Strain Indicators

Subtle rhombus and rhomboid parallelogram patterns on the Vaughan Lewis Glacier and the Gilkey Glacier (see Waag, 1975) apparently indicate strain orientations within the ice and overlying firn. The patterns in bubbly ice at the firn-ice interface, reflect stresses transferred from the ice into the firn, and form through differential recrystallization within narrow preferred zones.

Convergence of the Gilkey Glacier medial moraines indicates lateral shortening of approximately 30 % (Fig.55). The short axes of the rhombi and the obtuse angle bisectors of the rhomboids are parallel to the strike of extension crevasses, to the axis of shortening and thus to sigma 1. The long axes of the rhombi and the acute angle bisectors of the rhomboids are parallel to the foliation, to ice flow direction, and thus to sigma 3. The angles of the parallelograms are variable locally, but average 105 and 75 degrees, the variation probably reflecting intensity and duration of stress. Similar parallelograms occur within the troughs of wave bulges below the Vaughan Lewis Icefall (Fig. 53).

In the wave bulges, the foliation arcs parallel the wave and the long axes of the rhombi and acute angle bisectors of the rhomboids parallel the foliation, the direction of elongation and thus sigma 3. The short axes of the rhombi and the obtuse angle biscetors of the rhomboids parallel the strike of radial crevasses and sigma 1. (again see Waag, 1975)

Although the precise mechanisms and conditions of formation of the parallelograms are not yet understood, the geometry is curiously similar to quadramodal plots of c-axis orientations in some ice petrofabric studies. Detailed quantitative measurements of strain within both glaciers are planned, but strain data to date and the firn folds (Fig.73a) support the above stress analysis.

D. Morphology and Origin of Wave Bulges and Ogives

During the past six years, continuing measurements and surveys have been made of englacial structures in the apron sector below the Vaughan Lewis and Herbert Glacier Icefalls on the Juneau Icefield (see Fig.53 and 73b). Following the concept of Miller (1968), the rotated-foliation theory of wave-ogive and ogive formation is further elucidated, as opposed to (1) the ablation theory of Nye (1967) *; (2) the varying longitudinal stress theory (Freers, 1965) and (3) the compressional rotation theories discussed by Havas (1965), Kittrdge (1967), L.R. Miller (1970) and Dittrich (in this report and 1975). To some extent it is probable that elements of each concept pertain. The folded-foliation theory is given more credence by our recent studies, therefore its essence is reviewed with respect to the Vaughan Lewis Glacier with further details given in Miller, 1975a.

* personnel discussion with Dr. J.F. Nye at Camp 18, August, 1967.

Morphological Character of the Vaughan Lewis Glacier

The glacier is comprised of five glaciomorphic zones, noted as A through E in Figure 74. The prime nourishment zone (A) is the highest. A discharge zone (the icefall or Zone B) lies between 5000 and 3000 feet. Below this(p. 91a) is an expanded foot of the icefall forming an ice apron (C) and characterized by splaying radial crevasses. This is the wave-bulge zone. The next lower level is in the valley glacier sector, or the zone of maximum wastage (D). Lastly is the terminal zone where the glacier calves into an ice-dammed lake in Avalanche Canyon, just north of Camp 23 (Fig. 2).

The mean névé-line since 1965 has been at or above 3800 feet, hence above the base of the icefall and high enough to permit study of englacial structures. These structures are manifest at the glacier surface, on crevasse walls and in marginal collapse zones. They include transverse tension crevasses (Zones A and B); oblique and splaying crevasses and low-angled thrust faults (Zone C); tectonic (flow) foliation dramatically exposed at the surface (Zones C and D); normal and recumbent englacial folds plus surface wave-bulges with the foliated structures deforming into ogives (Zone C); and cross-cutting secondary ice veins and white-ice inlays exposed on the ice apron down-glacier (Zones C and D). To understand and explain them each of these fractures must be related to the total systems analysis of the structural glaciology of the glacier as a whole and cannot be considered as separate entities.

Wave Bulges and Ogives, a Differentiation

The sequence of upbulged surface waves and relative thickening in the apron region is considered to be the result of annual variations in longitudinal stress produced by regime changes in the névé (L. Miller, 1970). This is abetted by the impinging of ice against a bedrock threshold shown by seismic depth soundings taken in the 1965 JIRP field season (Kittredge, 1967). As the wave bulges pass over the threshold and the bedrock gradient increases, compressive stress is replaced by tension and the bulges rapidly attenuate.

The englacial ogive pattern so well displayed down-glacier is attributed to hyperbolic deformation of the folded tectonic foliation, further exposed by ablation. As the ogives pass down-valley in the two-dimensional expression they assume more parabolic form. Ogives are coincident with the bands in the wave-bulges, but in this up-glacier area they have no surface amplitude and show no detectable outlining with respect to the bulges. But the attenuation of bulges downstream increases the delineation of the ogives because more and more debris becomes concentrated at the surface from ablation of the basal foliation, thus concentrating dirty zones along banded lines. Some of the intervening white ice inlays are of clean and bubbly ice. Miller (1968, 1975a) considers these to represent compressed crevasse fillings from winter snow avalanches in Zone B at the base of the icefall. Waag (personal communication) suggests the possibility that they are mylonitic zones of intensified shear. Corroboration of Miller's idea, however, seems to come from the disappearance of these inlays far down-valley (Freers, 1966).

Variable deformation of the ogive patterns from the pointed arch (projectile) form to more rounded and parabolic forms is ascribed to streaming (parabolic) flow, which results when the glacier is undergoing minimal basal slippage. When such slippage increases in consequence of the rectilinear or plug-mode of flow, these arched bands even become rectangular in pattern, giving rise to the term rectilinear flow (Miller, 1973b). Variations of this kind are

found in the down-valley sector of the Vaughan Lewis Glacier (Fig. 70). For this the term "ogive" may not be fully appropriate, except in cases where there are geometrically true "ogive" forms. For more general usage, it has been suggested that these features may be called arch-bands (Miller, 1954).

For our purposes the term wave-bulges (wave-ogives) is reserved for those large surface undulations which begin to form at the base of icefalls (Fig. 53) and which in this case in the ice apron sector have their fullest development with amplitudes up to 80 feet and wave lengths of 300 or more feet. The general term ogive is still used, however, but is reserved for reference to arch-banded structures exposed two-dimensionally on the surface in Zone D after the wave-bulge affect has attenuated.

For convenience, the term wave-ogive has value if specifically referring to the 3-D banded structures revealed in the bulge zone. These bands relate to a remarkable series of tectonic folia (Fig. 73b). They are also well shown at the surface and margins of the apron below the icefall.

Folded Foliation Theory of Ogive Formation

Field studies between 1962 and present permit the further development of a unique theory of wave-ogive formation, first expressed at the Symposium on Surging Glaciers and their Geologic Effects in 1968 (Miller, 1968). The theory involves rapid down-valley movement of an icefall and its maximized longitudinal stress, which is responsible for development of recumbent isoclinal fold structures. Although rarely exposed at the surface, these have been found in rotated folia at the glacier's margin. The overturning is occasioned by a sudden change of declivity as ice moves over buried bedrock cliffs. The folds are subsequently righted into more normal positions as the glacier passes from a zone of tensile stress to a zone where the principal stress becomes compressive (Fig.73b).. The character of such foliation "folds" is accentuated at lower elevations where truncation by ablation exposes remnants of them in a way similar to erosion of plunging sedimentary folds in bedrock of the Appalachians.

The manner in which these folds are produced from basal foliation parallel to the glacier bed in Zone A, to deformed and overturned folia in Zone B and finally to re-orientated positions in Zone C, is depicted in Figure 73b. The relationship of alternating clean and dirty ice in the sector of maximum ogive exposure farther down-glacier may be a consequence of debris variability in the flow foliation.

The presence of low-angled thrust surfaces and of inlays of what have been suggested as compressed segments of young, bubbly glacier ice are diagnostic, as is the presence of splaying radial crevasses from extension effects in the ice as it emerges from the construction of its bedrock defile (bottom of Zone B). These are all expressions of tectonic processes in the lower icefall zone. The crux of the explanation lies in a plunge-pool effect due to existence of the bedrock threshold shown on the schematic profile in Zones B and C (Fig.s 73b,74). The bedrock topography provides just the right compression to upbulge the rapidly flowing ice before its surface is levelled by atmospheric ablation processes. Or indeed, as Dittrich suggests next in this report, a rotational form of flow in and below the icefall would produce the same type of effect.

E. Rotational Flow on the Vaughan Lewis Glacier

One of the more controversial postulates in glaciology is the concept of rotational flow (Lewis, 1949; McCall, 1953). In an effort to determine to what extent this mechanism could affect the development of wave-bulges (Fig. 53) in the Vaughan Lewis Glacier, approximate flow rates from above and below the Vaughan Lewis Icefall have been calculated from transverse velocity and seismic profiles in 1971-72. The surface features of the Vaughan Lewis Glacier and adjoining Gilkey and Watchamacallit Glacier are interpreted in terms of the surface velocity field (Fig. 75a). The degree to which the flow within the Vaughan Lewis Icefall is rotational is estimated by comparing the velocity of irrotational flow, as calculated from measured flow rates (Fig. 75b,c). The measured longitudinal velocity profile corresponds to highly rotational flow. This quantitative measurement might support earlier postulates that the formation process of wave-bulges and associated ogives to some extent relate to a rotational flow mechanism as well as compression in and below the icefall zone. Full details of this tantalizing concept are presented elsewhere by one of our research workers, W. Dittrich (1975).

F. Micro-Strain Measurements and the Question of Glacier Tides

Further to the foregoing, Dr. Gordon Warner (1972, 1973, 1974 *) has conducted research in micro-strain measurement on several Juneau Icefield glaciers using unbonded single-wire strain gages. The aim of this research was to develop a more precise strain-measuring method than the taping and survey instrument techniques heretofore available (e.g. Wu and Christianson, 1964) and one which was dependable under difficult field conditions.

The measuring technique which was developed uses unbonded electrical-resistance strain gages which consist of single strands of 5-mil Constantan wire 10-feet long. Six gages are embedded in the glacier-ice surface in the form of two delta rosettes in order to obtain strain at a point with some redundancy of data in this two-dimensional problem. The rigid-body rotation of the gage anchor posts was measured by sensitive inclinometers in order to assess the effect of pressure melting on the strain data. The data are interpreted using cross-correlation and best-fit programs to yield maximum shear-strain rate and average normal-strain rate.

Strain readings were conducted on the Ptarmigan Glacier. The maximum shear-strain rate at the surface ranged from 0.25 to 1.2×10^{-6}/hr., which agrees with estimates derived from known flow rates. The wire gages were found to adhere to the ice well enough to make the gage anchor posts unnecessary. Pressure melting is therefore insignificant. A tolerance of ± 6.0 microstrain was determined for the strain gages.

After two summers (1972, 1973) of measurements using this method, Dr. Warner has found a unique diurnal fluctuation in micro-stress which may not be meteorologically related. The suspicion is that we are detecting glacier tides similar to earth tides. The significance of this possibility is great in terms of actuating the movement of ice sheets. A bore-hole strain gage is currently being developed to alleviate the effects of daily meteorological changes in our continuing field research on this fundamental problem.

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* Also see Warner and Cloud, 1974

PART VIII PATTERNS OF RECENT GLACIER VARIATION AND NÉVÉ REGIMES ON THE JUNEAU ICEFIELD

Fundamental to the integrated research aims of this regional terrain study and the glacial processes which dominate is an appreciation of the morphology and present position of the termini of the main glaciers involved.* Of further import is an understanding of the former positions of these ice tongues as they relate to the evolution of relict landforms...moraines, cirques, berms and the like. Thus, brief notation is made of extensive reports recently published on these matters by members of our research group, including pertinent papers during the interval of this contract. Of special importance are Miller (1972, 1973, 1974, 1975); Egan (1971); Jones (1974; 1975); Tallman (1972, 1975) and Zenone (1972). Earlier reports providing further background on the regional glacier pulsations are by Beschel and Egan (1965); Egan (1971); Egan et al (1968); Konecny (1966); Lawrence (1950); Miller et al (1968) ; Chrzanowski (1968) and Miller (1964a, 1967, 1969 and 1971a).

The regional regime pattern on the Juneau Icefield over the past 30 years has been one of accentuated downwasting and retreat in terms of the number of glacier termini involved (some 32), with only five showing equilibrium positions and but two in a state of vigorous advance. Thus 80 per cent have dominantly negative regimes. However, in terms of the total area and volumes of ice involved, the 6 per cent (2 in number) in advance added to the 15 per cent (5 in number) in equilibrium represent more than 50 per cent of the total volume of glacial ice on the icefield. Thus the regime trend of the icefield over the past decade is healthy, and one that in the overall regional sense is producing a continuous build-up and enlargement of the total volume of ice in the presently glacierized area.

As discussed in Part II-F, this renewed growth of the main highland glaciers in this region is allied to a current 80-90 year climatic cycle, the cooling end of which will apparently bottom out about the year 2000. In consequence, the lower névés (below 4500 feet and down to 3000 feet) are experiencing renewed accumulation, whereas the upper névés (4500 feet and higher) are generally receiving less snow each year. Glacier variations in the Taku and Atlin Districts over the last 400 to 600 years of the Alaskan Little Ice Age have been well inventoried by Lawrence (1950) and Miller (1964, 1971 a and b, and 1975a).

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* Again refer to examples in photo supplement, especially page 91d.

PART IX PERIGLACIAL INVESTIGATIONS -LANDFORMS AND PROCESSES

A. Research on Tank and Tor Topography

As discussed in Parts X to XII, the limits of middle and upper Wisconsinan glaciation can often be identified by the presence of large tors (bedrock knobs) and associated tanks (rock basins) or bedrock depressions (p. 91). These abound at elevations higher than 4000 feet (1200 m) at the northern end of the Atlin valley and above 5000 feet (1500 m) at the south end. This is because their evolution has demanded a prolonged period of subaerial exposure.

Similarly, on the peripheral ridges of the main Icefield in coastal Alaska Wisconsinan limits are well shown by such topographic expression. One such locale is the 4000-foot rock ridge separating Ptarmigan Glacier from the Canyon Creek cirque system (Fig. 23). Here a unique development (p. 91b) of small-magnitude tors and tanks characterizes this deglaciated ridge. Special study of these features was conducted in 1972 and 1973 (Zwick et al, 1974).

The research locale was a ridge zone about one mile long and one quarter mile wide. Width of the tors ranged from several meters to 20 meters, with lengths of 40 meters and heights and depths of 1 to 5 meters above and below the immediate surrounding surface. The tanks have served as nivation hollows for localized firn-packs during the most recent decade of cooling climate, with the firn in some being retained over several summers. The tanks are related to selective plucking between granitized lithologies (gneisses and migmatites) and the more schistose units exposed on these ridges. Their geometry is controlled by relict bedding structures and joint sets. The major tank orientation is N 45°W to N 60°W, roughly parallel to the N 47°W direction of ice movement in maximum Wisconsinan time. "Bedding" structures on these granitized units also strike N 50°W ± 10°, with major joint sets of N 55°W ± 20° and N 45° E± 15°.

Associated with the tank and tor relief are single and compound solifluction lobes which lie as terraces on sides of the crestal ridges. Distinct morphological characteristics provide measurable aspects regarding relative age, nature of movement and association of dimension, slope, lobe thickness and size of clasts. Some information was gleaned on recent sedimentation and organic development rates in the tanks. A quantitative consideration of all factors was attempted. One conclusion is that the stability of detrital mantle on these frost-riven slopes is greatest when there is a combination of gentle slope and coarse material. Fabric studies suggest that variations in width/length ratios and slope have little effect on the nature of deformation during mass wastage (Fleisher, 1972).

The summit relief along these ridges is controlled by the abundant tors and tank basins, with the tors often found to stand as stone walls along the crest and to be located along the strike of the more felsic facies in the granite gneiss. Solifluction lobes a few meters across and 4-8 meters long typically form at the base of the tors.

Our investigations have shown that (1) tors are weathering remnants of middle to late Wisconsinan time; (2) many small tors are presently forming and contributing material to the flanking solifluction lobes; (3) the tanks are presumed to be a result of glacial quarrying by early Wisconsinan ice which overrode the arêtes; and (4) because of the elevation of the ridge crests,

the overriding ice could not relate to present glacial drainages. Most probably its provenence was coastward flowage of Taku Valley ice from the interior of the Northern Boundary Range during maximum early Wisconsinan glaciation.

B. Patterned Ground

At an elevation of 2500 feet (800m) at the north end of the Camp 17A arête (see foregoing), an array of large relict stone circles has been studied (v.p. 91b). The circles are 2 to 5 meters across with thick and well-developed heath mattes filling their centers. The location of these circles, in positions just outside of the sequence of Neoglacial moraines noted in the discussion of the Ptarmigan Glacier, Part XA, reveals that at the time of the ring formation excessive frost conditions existed some 1500 feet below the zone of maximum tor and tank development. This supports the idea that these striking periglacial features evolved during the waning phase of the most recent maximum glaciation just prior to the pronounced warming of the Holocene. During the Thermal Maximum, there was minimum frost action, but since the onset of Neoglacial conditions re-intensified frost weathering and active solifluction have been the prime periglacial processes affecting these slopes.

A large number of active stone rings and associated stripes and solifluction lobes has been investigated in the Camp 29 sector of the Cathedral Glacier massif (Fig. 3) as well as on higher ridges and tundra plateaux above the 5500-foot permafrost level in the Atlin region. Relict stone circles are found here at elevations down to 3000 feet. Reports on this research are forthcoming (e.g., Jones, 1975; Nelson, 1975). Active patterned ground and felsenmeers associated with permafrost have also been studied on high ridges in the central Juneau Icefield area, especially near Camps 8 and 25 (i.e., above 7000 feet elevation).

C. Research on the Age of Palsas, Upland Peat Plateaux and Rock Glaciers

The discovery of a number of palsas (hydrolaccoliths) and ice-cored peat plateaux in high-level swamps of the Atlin tundra region (Miller, 1973) has opened up the opportunity for some unique research on terrain of transitional permafrost character. The best developed palsas are on the flattish upland (v. p. 91a) at the head of valleys above 3000 feet elevation, up to the permafrost seep level at 5500 feet, above which elevation true permafrost structures occur. Of particular interest were the palsas in the intermediate and upper Fourth of July Creek Valley some 20 miles northeast of Atlin (Fig. 3). Here the palsas are quite similar to those described in Finalnd by one of our researchers, Dr. Matti Seppala (1972). The Fourth of July Creek palsas lie in the midst of a striking swarm of late Wisconsinan eskers (Tallman, 1975) and have been investigated by electrical resistivity means (Tallman, 1972 and geophysics section of this report).

Of particular interest here is our research on the age and evolution of these features, the significance of which lies in their sensitivity to climatic change (Miller, 1975b). A trench was dug through one large peat plateau at 3300 feet elevation and a collection of twigs collected from the frozen silt-clay interface at its base (i.e., the stratigraphically lowest zone of organics). A radiocarbon date of 9315 ± 540 C-14 years B.P. (Geochron 1972, Sample GX 2694) indicates that this is essentially early Holocene vegetation and hence that the area began to be deglaciated 9 to 10,000 years B.P. Similar dates from

a reactivated palsa in the same locale gave a date of approximately 8050 ± 430 C-14 years B.P. (Geochron, 1972, sample GX 2695). A sample from near the base of a palsa bog at 3000 feet elevation gave 3,090 ± 170 C-14 years B.P., which indicates this palsa developed in Neoglacial time. Zones of "dead" palsas have also been identified, suggesting a growth and decay periodicity which may relate to a periodic growth and self-destruction that is now entirely climatically controlled.

When allied with C-14 dates in other bogs in the Atlin valley (Anderson, 1970; Miller and Anderson, 1974), this information suggests that to some extent the palsas are thermally transitional frost-related features, sensitive to vertical shifts in the regional freezing level and more importantly to changes in duration of the winter season. Thus they may provide geomorphic evidence for secular climatic trends well outside of the present glacial position. The question of climatic connotation remains sufficiently uncertain, however, that our research on this phenomenon is proceeding. Further studies in 1975 extended into a region of more extensive palsa development in the Ross River area, some 500 miles to the north. This will be reported on in a later publication.

Our rock glacier research in the Atlin sector also has implications regarding Holocene climates, as the rock glaciers all truncate late-Glacial moraines. The rock glaciers are usually associated with cirque glacier (alpine stage) development in the early Holocene, following deglaciation of the main Atlin valley. Key aspects of this research are noted in the section below.

D. Tundra Terrain and the Permafrost Seep Level

In the Atlin area, permanently frozen ground is found above the elevations of 5500 feet and on the Juneau Icefield above 6000 feet. Below these levels there is a summer-long seepage of melt-water out of the ground ice. Thus tundra plateaux at this elevation in the highland areas bordering the Atlin valley are often saturated with water even though the regional climate is semi-arid (10 to 12 inches of w.e. precipitation per year).

In this zone, stone stripes and solifluction lobes abound and at elevations above 6000 feet large and active stone rings and polygons occur, some with dimensions of 4 to 5 meters across. Many relict stone circles are also found at elevations down to 3000 feet, revealing the effects of Pleistocene and early Holocene climatic change. The region is propitious for further research on a type of terrain that is found in many areas of the Yukon and Alaska near and beyond the Arctic Circle.

E. GEOBOTANICAL AND SOILS STUDIES

In the Atlin region and adjoining sectors of the Yukon, rock glaciers have been propitious for lichenometric study (e.g., Miller and Anderson, 1968). In the Atlin area these studies and our soil vegetation research are still in progress. Preliminary results have been presented by Anderson (1970 and Miller and Anderson, 1974); Kostoris (1973)*; Tomlinson (1973); See (1974); Jones (1975); Miller (1975a); Tallman (1975) and Magnuson (1975)*. The approach used in the lichenometric studies being carried out on these features and the moraines of the Junaau Icefield glaciers is discussed in the following section, in which is considered the example of the Ptarmigan Glacier near Juneau.

Exhumed bog levels in the Boulder Creek sector of the Atlin region have yielded some significant results (Miller and Tallman, 1975). A soil and soil-vegetation study has also been initiated by Lietzke (1975) to extend earlier work on soils and paleosols of nunataks on the Juneau Icefield (Lietzke and Whiteside, 1972). Spodosol development on the youngest till and outwash drift in the Atlin valley has been shown to have strong similarity with the old podsol soils described in the literature of glacial soils in the northern Great Lakes region.

A regional taxonomic study of the vegetation in the Atlin area is also being conducted by Dr. J.H. Anderson of the Institute of Arctic Biology, University of Alaska (Anderson, 1970 and manuscript in preparation).

Further taxonomic research on the plants and soil relationship of Birch Mountain, Teresa Island, near Atlin (Fig. 3) is being conducted by Steven Buttrick of the Department of Botany, University of British Columbia (Buttrick, 1974, 1975)*. Mr. Buttrick's research includes investigation of the tundra area at and above timberline, with some attention to the relationship of relict patterned ground.

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* In open file progress reports on 1973-75 field work, JIRP, Found. for Glacier and Environmental Research, Seattle.

Part X QUATERNARY TERRAIN FEATURES

This terrain study provides a basis for outlining the Quaternary history of this key region at the northern end of the Pleistocene Cordilleran Ice Sheet. In the coastal sector, dense forest cover and mass wastage make landform analyses and interpretations of the geologic history difficult, but by extending the studies inland across the Icefield, as well as to the upper Taku River Valley and in the Atlin region (Fig. 3) on the continental flank of the Boundary Range, we can draw useful comparisons from less afforested areas in the Canadian sector. Recourse is made also to the evolution of erosional topography with the phases of glaciation suggested by study of the morphology and spacing of tandem cirques and rock-shouldered berms, as well as relict periglacial features found in coastal and inland areas. First considered are the surface deposits in the Taku District near Juneau. During the course of this, reference will be made to the chronology chart, Table XV.

A. THE TAKU DISTRICT

Tills and Diamictons

A fossil-rich drift sheet partially of submarine origin was investigated in the Juneau area (Fig. 1). At the top are terrestrial sand and gravel deposits from tributary valleys that grade in places into massive slump areas. This work extends earlier mapping of these deposits (Miller, 1956, 1963) and reveals that continuing mass wastage has taken place in Neoglacial time. Radiocarbon dating of peat overlying the uppermost member of this formation proves it to be late Pleistocene and early Holocene in age...ca., 8,500 years B.P. (University of Alaska, Radiocarbon Dating Lab., 1974). Carbon isotope dating of an avalanche-sheared and colluvium-covered stump (ca 800 C-14 years B.P.) suggests quite recent effects in the topmost layer possibly by earthquake if not climatic-induced debris avalanches in the 12th century.

A geologic appraisal of the formation as part of an earthquake hazard study of Alaskan coastal communities (R.D. Miller, 1973) has considered these glacial tills as diamicton (slump) deposits contaminated throughout by berg-rafted boulders. Although differences of opinion may rest on how the term diamicton is used, we suggest that the Gastineau Formation be referred to as a slumped, glacio-marine till-sheet with its submarine glacial origin emphasized. Only the upper surface zone would be affected by berg-rafting, i.e. from the final phase of deglaciation in a subaqueous environment. It is recognized, of course, that the drift sheet has been subjected to varying degrees of flow deformation well after deglaciation, which process is presumed to be largely responsible for the fracturing and further dissemination of embedded fossil shells.

A two-phased nature of the glacio-marine facies is accentuated by differences in weathering on the two main units involved. The weathering is a function of both age and environment (i.e., reducing vs oxidizing conditions). Of special interest is that the lower blue-gray unit extends beneath Neoglacial moraines of the Mendenhall Valley up to an elevation of 150 feet (50 m) above mean sea level. Also the oxidized upper unit is found in fossil-rich zones at elevations of 300 to 500 feet (90-150 m) on the sides of gulleys and valley tributaries in the Gastineau Channel and Lynn Canal sector (Miller, 1956). Thus at least 500 feet (150 m) of sea-level change has taken place since the advent of the Holocene. This is

corroborated by Twenhofel (1952) who, on epierogenic evidence, suggested that at least half of the shoreline changes in this sector of Alaska have been due to post-Glacial rebound.

Cirques and Berms

One of the most striking geomorphic features of the coastal ridge flanks of the Boundary Range is a magnificent array of well-developed and for the most part ice-abandoned cirques, many in tandem sequences. Regardless of structural or lithologic character of the bedrock, in the Alaskan sector the cirques form a five-fold sequence (C-I to C-5) up to 4,000 feet elevation with a roughly 700-foot spacing in elevation zones from 300 to 3,200 feet (Miller, 1961). Above the 3,200-foot level there are four additional cirque systems (C-6 to C-9), but these are largely ice-filled, with less distinct spacing, and lie between 4,000 and 6,200 feet.

On the continental flank of the range, five distinct cirque levels are identified between 4,000 and 6,500 feet. The upper four levels are roughly parallel to the elevation distribution of the higher clearly identifiable cirque systems on the Juneau Icefield. The comparative distribution reflects a pronounced rise of Pleistocene snowlines inland (Squyres, Miller and Jones, 1974).

In Table XI, the mean elevations are given for each cirque system, designated C-I to C-9. Comparison is also made with interpretations from the Alexander Archipelago in southeastern Alaska (Swanston, 1967) and in the Atlin, B.C. area on the Cathedral Massif (Jones, 1975) and the Fourth of July Creek Valley (Tallman, 1975). Allied with the cirque distribution is an equally remarkable sequence of rock-shouldered berms along the walls of main trunk valleys in the region (Miller, 1963). These are reflections of major phases of valley glaciation and reveal a five-fold tandem pattern with cyclopean stairs similar to those in the tandem cirques. In any direct attempt to correlate these berm levels with the cirque sequence, the upper berms, of course, integrate with the lower levels of cirques (v. Table XIII).

Significance of the cirque and berm sequence lies in an apparently close relationship between cirque floor elevations and mean freezing levels (elevations of maximum snowfall) during sequential stages of the Wisconsinan. The main development of cirques is attributed to waxing and waning phases, abetted by periglacial processes in intra-glacial phases (Miller, 1961).

Reference (Table XIII) is also made to the glacial stage most recently affecting each of the cirque and berm levels (discussed further on page 124).

Early Holocene Periglacial Conditions

Relict stone circles of early Holocene age are found at 2,500 to 2,800 feet elevation on benches and shoulders of the coastal valleys in southeastern Alaska as well as at slightly higher elevations in the interior Atlin region. These are decimated structures, as today true periglacial conditions occur only above 4,000 and 5,500 feet respectively in each sector. Two examples from the Camp 17 area near Juneau and the upper Fourth of July Creek Valley in the Atlin region are found in Appendix H.

TABLE XII

COMPARISON OF MEAN CIRQUE ELEVATIONS IN THE BOUNDARY RANGE SHOWING CORRELATION WITH RESPECT TO COASTAL AND INLAND SECTORS

	SOUTHEASTERN ALASKA FLANK OF RANGE		CANADIAN CONTINENTAL FLANK OF RANGE	
Cirque Level	Juneau Icefield and The Taku District* (Miller, 1956, 1961)	Prince of Wales Island (Swanston, 1967)	Fourth of July Creek Region (Tallman, 1975)	Cathedral Range (Jones, 1975) (Miller, 1975)
C1	300 ft. (90m)	0-500 ft. (0-150m)	———	———
C2	1000 ft. (305m)	650-950 ft. ([illegible]0-290m)	———	———
C3	1800 ft. (550m) King Salmon-Port Huron?	1050-1350 ft. (320-410m)	4100 ft. (1249m)	———
C4	2500 ft. (760m) Tulsequah-Early Valders?	1450-1950 ft. (440-590m)	4900 ft. (1494m)	4500 ft. (1370m)
C5	3200 ft. (975m) Sittakany-Late Valders?	2050-2650 ft. (625-805m)	5500 ft. (1676m)	5100 ft. (1550m)
C6	3900 ft. (1190m) Early Holocene and Neo-glacial, including today		6000 ft. (1829m)	5800 ft. (1770m)
C7	4600 ft. (1400m) Thermal Maximum			6500 ft. (1980m)
C8	5500 ft. (1676m) Thermal Maximum and Sangamonian?			
C9	6200 ft. Thermal Maximum and Sangamonian			

* with postulated reference to these glacial stages most recently filling these cirques with ice

Development of this patterned ground relates to exposure of the ground surface in the final retrogressive phase of the Wisconsinan and probably also to early Holocene climatic conditions. This is based on C-14 samples of basal organic horizons in bogs where some of these features are found, giving deglaciation at about 9,700 C-14 years B.P. (Geochron, 1973).

The abundance of tanks and tors on maritime ridges of the Coast Range at 3,000 to 4,000 feet and in the Atlin region at 4,500 to 5,500 feet is further testimony to the long enduring intensity of frost climates in late Glacial and early Holocene time (Fleisher, 1972; Zwick et al, 1974; Tallman, 1975). Along Cairn Ridge near Camp 17 (Fig. 23), numerous small-sized stone rings (up to 0.5 meter) are found. Most of these are relict but a few have reactivated centers. At higher elevations on the central nunataks of the Juneau Icefield and at levels greater than 5,000 feet in the Atlin sector, stone circles, stone stripes, and other evidences of relict and presently developing patterned ground are found in abundance.

Late Holocene Moraines

Our investigation of the intermediate-elevation valley of the Ptarmigan Glacier six miles east of the Juneau Airport has revealed a sequence of late-Glacial and Holocene moraines and allied kame and outwash features that are typical of the maritime sector. There are seven distinct moraines, the first of which is pre-Holocene and characterized by stabilized felsenmeers between zones of thick heather cover west of Camp 17A.

The four most recent terminal moraines are untruncated and lie in the inner Ptarmigan Valley (Table XIV). Lichenometric measurements give the best time-stratigraphic information to supplement the morpho-stratigraphic details. This was accomplished by statistical sampling of hundreds of thalli diameters of crustose lichen on each moraine using Placopsis gelida (growth rate of 1 cm per 25 years) and Rhyzocarpon geographicum (estimated growth rate of 1 cm per 80-100 years) (See, 1974).

In this sequence, the time-span between moraines I, II and III appeared to fit the 80-90 year cyclicity described for this region (Miller, 1969, 1971, 1973,1974). The time spacing between moraines III and IV appears to represent only a 20-30 year pulsation. The present ice position is slowly downwasting at the first narrowing of the inner valley. This moraine sequence is significant as a base for further comparison with glacier regime changes in other valleys of the Boundary Range.

Pleistocene Erosional Landforms and Morphogenetic Phases of Glaciation

The character and significance of erosional sequences given by berms and cirques have been further discussed in several recent theses and in other publications (Egan, 1971; Jones, 1975; Miller, 1961, 1963; Tallman, 1975). A description of the morphological nature of the sequential phases of glaciation pertaining to the Cordilleran Pleistocene has been described elsewhere from field evidences in this area of the Boundary Range (Miller, 1964b). The morphogenetic character of these glacial phases and their relative magnitudes are noted in Table XIII.

TABLE XIII

TYPES OF GLACIATION DURING THE WISCONSINAN ACCORDING TO ELEVATION OF CIRQUES AND BERMS WITH SUGGESTED CHRONOLOGIC RELATIONSHIPS

Index Cirques (with elevation in feet)	Berm Levels in Main Valleys	Nature and Relative Magnitude of Regional Glaciation*	Suggested Chronology with Short Designations	Probable Time Range of Development
C6-3500**	B1	Retracted Ice-field	Neoglacial	
C5-3200	B2	Extended Ice-field		GIVc C5
C4-2500	B3	Lesser Mountain Ice-sheet	Late Wisconsinan	GIVb C4
C3-1800	B4	Lesser Mountain Ice-sheet		GIVa C3
C2-1100	B5	Intermediate Mountain Ice-sheet	Upper Middle Wisconsinan	GIII C2
C1-(embayment stage)	B6	Greater Mountain Ice-sheet	Lower Middle Wisconsinan	GII C1
C1-350 (initiation stage)				
Over-deepening of longitudinal valley into present U-form		Greater Mountain Ice-sheet	Early Wisconsinan	GI
Cross erosion of upper slopes and ridges, deepening of main valleys		Intermontane Ice Cap	Pre-Wisconsinan	

* As related to major excavation of cirques at reference level

** Level of present semi-permanent névé-line on western side of icefield

TABLE XIV

LITTLE ICE AGE MORAINES IN PTARMIGAN VALLEY, JUNEAU ICEFIELD

	Geomorphic Features	Lichenometric Characteristics	Estimated Date of Formation
M1.	Outermost Little Ice Age Moraine: A bold terminal arcuate in form; 5m high and 20 m wide, with huge boulders in consolidated till. Many lichens.	Considerable heath, grass, and sedge matte. Large number of *Rhysocarpon* thalli of 2 to 2.5cm mean diameter.	ca. 1715-1760
M2.	An arcuate, more subdued, and less vegetated moraine; low hummocks 3m in height; rooted willows of 2cms diam. Large boulder fragments; Intermediate development of lichens.	Some sedge and grass, many *Rhizocarpon* of 1.5cm mean diameter. A few *Placopsis* of 2-2.5cm.	ca. 1820-1830
M3.	Only slightly vegetated moraine, subdued in relief. Some slump zones, with coarser materials than older moraines. Few lichens.	Slight growth of grass and moss. No *Rhysocarpon*. *Placopsis* up to 2.5cm.	ca. 1900-1905
M4.	Fresh-appearing unslumped hummocks, essentially non-vegetated ground moraine. Very few lichens.	Only a few *Placopsis*, mean diameters, 1-2cm.	ca. 1925-1930
M5.	Ground moraine near present still-stand of Ptarmigan Glacier terminus. Relatively fresh till with unslumped hummocks. No lichens.	Fresh-appearing material; no vegetation.	ca. 1960-present

Based on the cirque and berm sequence, the elevations of index cirques and the relative position of main valley berms on the coast and in the interior as noted in Table XIII presumably relate to the nature and magnitude of these regional phases of glaciation. Because the cirques and berms are the product of multiple erosive cycles during Wisconsinan time, this table postulates a reasonable period of development through a number of glacial stages. The relative lengths of the arrows do not represent time magnitudes but only suggested chronologic ranges.

Pleistocene Depositional Stages

The distinction between stages and phases is that stages have time-stratigraphic meaning, whereas phases represent glaciation magnitudes; i.e., it is a geometric concept and hence only has morphostratigraphic significance. A phase of glaciation, of course, may embrace repeated stages.

A number of Pleistocene stages have been identified in the overall region of our concern, but the relationships are made complex by multiple provenances of ice. Both in the coastal Taku District and in the interior Atlin region, two widespread occurrences of middle to late-Wisconsinan tills are reported, but these are not to be construed as contemporaneous. Regionally, however, there is evidence of a widespread drift sheet attributed to the early Wisconsinan. Pre-Wisconsinan glaciation too was all-inclusive, but the only evidence is high-level U-shaped cols and formerly rounded summits, today seen as severely serrated ridges due to the intensity of Wisconsinan erosional processes. Also, even though early Wisconsinan glaciation left extensive erratics on ridges above later glaciation limits, much of the evidence has been destroyed by the intensity of subsequent periglacial processes. A factor in this is the generally cold climatic condition at high elevations that even during intraglacial intervals retarded soil development. The presence of weathered and relatively deep till soils postdating the early Wisconsinan maximum suggests that this glaciation was prior to 50,000 years B.P.

Surface Stratigraphy on the Coast

The two youngest tills are found most commonly in the lower parts of valleys and on the coast in emerged fiord deposits (Miller, 1963; Swanston, 1967). In the Juneau area the top member is an indurated blue-gray boulder clay diamicton with coarse clastics, designated in this sector as the upper segment of the Gastineau Formation. Because of the well distributed boulder and cobble content and much intercalated outwash material, it may be essentially subaqueous combined with glacial in origin (see earlier discussion). Our studies reveal it to contain late-Wisconsinan Clinocardium sp shells and to rest on an older till with zones of mixed colluvium and glaciofluvial facies. The lower segment is a more definite compact and ill-sorted till with a mild weathering profile. In part, however, it too is a diamicton with included Leda-type shells. At higher levels it appears to correlate with a weathered silty facies containing a few large boulders lying stratigraphically below but often topographically above stained deltaic gravels considered to be very late-Wisconsinan in age. A number of such deltas occur at the mouths of distributory streams from the Coast Range icefields.

Above the boulder silt, to elevations of 520 feet (160 m), is a marine clay containing shells of Clincardium, Pectin and Maccma calcarea gemlin species. Many

of these have been disrupted and broken by subaqueous flow. What is apparently a counterpart of this two-phased sequence has been found in the Alexander Archipelago (Swanston, 1969). The weathering profile on the basal unit in this drift and an apparent correlate in the Atlin region appear to reflect intra-glacial conditions. This sequence, when taken as glacially-related and in conjunction with other evidences (such as the regional array of cirques and berms) is aiding in the gradual resolution of Wisconsinan chronology for the Taku District as presented in Table XV.

B. THE ATLIN DISTRICT

Inland Counterparts in the B.C.-Yukon

A probable inland correlate of the upper member of the Gastineau Formation is found in the Tulsequah and Atlin Lake regions (Fig. 3). In the latter area, the two surface tills have been well exposed by sluice operations in the main gold placer streams in the vicinity of Atlin, especially in Spruce, Pine and Boulder Creeks (Miller, 1975a; Tallman, 1975). In McKee Creek three till units occur. Radiocarbon dates on the section in Boulder Creek valley show that warmer and wetter conditions persisted in the interior prior to 40,000 years B.P. Other C-14 dates from bogs near the Alaska and Atlin Highways and in creek banks reveal that deglaciation in the Atlin region was well underway before 10,000 years B.P. (Anderson, 1970; Miller, 1975a; Tallman, 1975).

The second oldest till in the interior region (e.g., the basal one at Boulder and Pine Creeks and the intermediate one on McKee Creek) is much more altered by weathering than the lower till-like member of the Gastineau Formation. As it is in a much drier interior climate, this weathered till is considered to be older, leaving the lower Gastineau member to correlate with the upper McKee Creek unit (i.e., the youngest Atlin till). There appears, however, to be a genetic relationship between the lower Pine and Boulder Creek deposits and oxidized high-level moraines near timberline, and possibly to a paleosol horizon on higher nunataks of the Boundary Range (Lietzke and Whiteside, 1972). Intra-glacial conditions seem to be suggested (provisionally termed the Pine Creek Intraglacial in Table XV. There is correlation, too, with a two-fold Wisconsinan moraine and kame-moraine complex in the intermediate and upper Fourth of July Creek Valley northeast of Atlin (Tallman, 1975). Here northeasterly flowing ice lobes from the Tagish and Atlin Valleys flowed into a junction (interlobate) zone not far from the upper limit of northwesterly flowing ice from the Teslin and Gladys Lake Depression and via these channels from the Yukon Plateau west of the Cassiar Mountains.

There is also evidence in McKee Creek and in the high-elevation tundra plateaux of the Atlin region that a more intensive glaciation deposited an extensive earlier till(s). Research is proceeding on this significant aspect of what we believe to represent an early Wisconsinan glaciation. The complex nature of the regional glaciation is also recorded in a series of glacial moraines and fluvial and lacustrine terraces at high elevation in the Tagish, Atlin and Teslin Lakes area, as well as by the sequence of interior cirques noted in Table XII.

In the Atlin Lake region the final diminishment and retreat of Wisconsinan ice are documented by massive embankment moraines and pro-glacial valley-mouth deltas in the northern part of the main valley, as well as by swarms of esker

complexes in highland tributary valleys to the northeast (v. p. 91a). Carbon 14 dates reveal that ice remained on high plateau areas probably as recently as 8,000 years B.P. and that even the lower level ice did not melt away much prior to 10,000 years B.P.

Terrain Features of the Holocene

Our palynological and periglacial research provides an interpretation of Holocene glacio-climatic events in the Boundary Range. Aspects of the past 10,000 years have been discussed with respect to the Alaskan coast. In the interior, an array of relict palsas (hydrolaccoliths) in the Fourth of July Creek Valley provide especially useful information. Some of these are redeveloping in the present cooling trend (Miller, 1973a; Tallman, 1975). (Again v. p. 91a of this report.)

Key questions to ask are when did the Holocene begin and did it actually end at the beginning of Neoglacial time? The palynological records reported in Miller and Anderson (1974a) indicate that for the Neoglacial in the Boundary Range climatic cooling was well under way by about 2,500 years B.P. This is suggested also by a C-14 date from post-Glacial bogs in the Surprise Lake area (Geochron, 1974; Tallman, 1975), where significant climatic change appears to have begun about 2,770 C-14 years B.P. Evidence from the Mendenhall Glacier area, previously described, suggests that 2000-2200 years B.P. ice was overriding forest beds in coastal valleys. If we assume a 200 to 500-year buildup of glaciers in the Coast Range following the Thermal Maximum, a date of about 3,000 years B.P. is reasonable as the effective start of Neoglacial time in this sector of the North Pacific Cordillera.

A two-phased nature of the Neoglacial is suggested by the warm interval that produced a retraction about 1,100 to 700 years B.P. or from A.D. 900 to 1300 (Tallman, 1975; Egan, 1971). The best evidences are the C-14 dates on over-ridden trees in the Taku Glacier and Davidson Glacier areas (Miller, 1973a). Thus, the Alaskan Little Ice Age is assumed to have begun following this warm interval... i.e., approximately in the 1300's...a suggestion corroborated by the dendrocrhonology dates obtained on the Bucher Glacier moraines.

As for the early Holocene, samples of basal peat from palsa bogs in the Upper Fourth of July Creek Valley give several C-14 dates around 9,700 $\pm$ 540 years B.P. (Geochron, 1972). These peats represent the initiation of organic growth following deglaciation. This conclusion is further corroborated by the 10,000 years B.P. minimum date of ice recession from the lower end of Atlin Lake. Added to this is a C-14 date of 7812 $\pm$ 120 years B.P. for the basal peat horizon immediately above the younger diamicton till in the Gastineau Formation near Juneau (University of Alaska, Radiocarbon Lab., 1975), as well as a range of 9,000 to 11,000 years B.P. on peat and logs from the base of muskeg in the Lemon Creek and Montana Creek valleys near Lynn Canal (Heusser, 1960; Miller, 1975b).

Further testifying to the initiation of the Holocene right after the final retreat of late-Wisconsinan glaciers from the Atlin region is the truncation of intermediate elevation lateral moraines produced by Atlin Valley ice along the 3,000-foot contour of Atlin Mountain. Here rock glaciers from abandoned cirques at 5,000 to 6,000 feet have flowed down to truncate these moraines, proving that these periglacial features are less than 8,000 years old. A similar situation pertains in the rock glacier valley on the Cathedral Mountain Massif (Fig. 8b).

In view of all this, the Holocene climatic trends show a rather continuous warming from about 9,500 years B.P. to about 6,000 years B.P. This led up to the Thermal Maximum interval when mean annual temperatures in the interior region reached some 2°F (3°C) warmer than present. During this warm interval, the most intensive of periglacial processes also migrated to higher elevations and the much retracted Juneau and Stikine Icefields received a new input of accumulation on their crestal névés, as did the cirques at the elevation of levels 5 and 6. Today these cirques are some 700 feet higher than the lowermost cirque at present filled with ice in the interior Atlin region. They are also an equivalent elevation above today's permafrost seep level in these interior mountains.

Neoglaciation and The Alaskan Little Ice Age

Near the end of the Thermal Maximum, the Arctic Front (defined in Miller, 1956, 1973a) held a mean position some distance west of the inner reaches of the Inside Passage in the Alaskan Panhandle. Then wetter conditions and increased storminess prevailed on the continental flank of the Boundary Range.

The Thermal Maximum was followed by a significant cooling trend as the Arctic Front again moved well inland. This was coincident with decreased storminess and drier conditions with general cooling on the continental flank of the Range. Concurrently on the coast, relatively cooler and wetter conditions dominated and led to increased glaciation in maritime sectors of the icefield highland. The nature of these out-of-phase climatic trends has been documented and explained by Miller and Anderson in recent publications (1974a, 1974b).

It is now apparent that the time interval from about 3,000 to 900 years B.P. was as much as 4° F (2°C) cooler than today. This was an interval of intense glacial activity in the high center of the Boundary Range, and also it represented intense periglacial activity in the area of palsas and sporadic permafrost at the 3,000-3,500 foot level in the Atlin region. Then came a short and warmer interstadial persisting until about the end of the 13th century (Miller, Egan and Beschel, 1968). Since the 14th century the Boundary Range and its peripheral sectors have experienced relatively cooler and wetter conditions compared to the short preceding warm interval in the Middle Ages.

PART XI WISCONSINAN AND HOLOCENE CHRONOLOGY FOR THE NORTHERN BOUNDARY RANGE*

The array of glacial and glacio-fluvial landforms investigated in this large region at the northern end of the Pleistocene Cordilleran Ice Sheet permits a general chronology for the Quaternary Period to be developed. Although some interpretations are provisional, the chronology is tabulated in Table XV (Tallman, 1975; Miller and Tallman, 1975). It should be considered as a basis for future reference and a spur to further research, especially since in this region only evidence for Wisconsinan glaciation is certain.

On the teleconnectional assumption that major glacial phases represent major changes in global climate, the chronology and terminologies developed for the Taku and Atlin regions are, in order of sequence, compared with chronologies suggested by other workers in the adjoining regions of Alaska, Yukon and British Columbia (Karlstrom, 1964; Péwé et al, 1965; Péwé, 1975; Denton and Stuiver, 1967; Bostock, 1966; Hughes et al, 1969; Rutter, 1975). Comparison is also made with other key sectors of the northern tier of states that have been affected by similar rigors of climatic change and major changes of glacier position during the Pleistocene Epoch (Madole, Mahaney and Fahey, 1975; Easterbrook, 1963, 1975; Crandell, 1965; Flint, 1971; Terrasmae and Dreimanis, 1975; Black, 1975; Frye and Willman, 1973).

A. THE WISCONSINAN

Although the evidences cited are mainly morpho-erosional and morpho-stratigraphic, they are abetted by critical time-stratigraphic and soil-stratigraphic information. In the Taku District, the Wisconsinan Age is sub-divided into four glacio-climatic stages (sub-stages). These are an Early Juneau/Atlin Stage (early or pre-classical Wisconsinan) which was represented by a Greater Mountain Ice-sheet type of glaciation followed by a distinct and relatively cool intraglacial interval. Then the rebirth of glaciation reached the levels of a subsequent Greater Mountain Ice-sheet phase, finally culminating in the Gastineau/Sloko Stage, that lasted from perhaps 40,000 to 14,000 years B.P. Following this was a short intraglacial and then the Douglas/Inklin and Salmon Creek/Zohini Stages. With respect to each event, place names are cited referencing diagnostic features and type stratigraphic units (Miller, 1975a).

The latter two glacial events were represented by Intermediate and Lesser Mountain Ice-sheet phases of glaciation and on the basis of sequence should be equivalents of the Port Huron and Valders glaciations of the mid-continent chronology. In the Boundary Range, the Salmon Creek/Zohini Stage has at least three distinguishable pulsations termed the King Salmon, Tulsequah, and Sittakanay substages. After this came the high-cirque glaciation (C-4 to C-5 levels) at the beginning of the Holocene, with pulsations probably comparable to those of the Little Ice Age during the past few centuries.

In the Atlin region there are strong correlates of this Taku District chronology. Although a similar sequence pertains, there are some differences that require explanation. Here too, a pre-classical Wisconsinan (early Wisconsinan) stage is recognized, with evidence that its thermophysical character was probably Polar (Miller, 1975b; 1975c). Called a Pre-Atlin I Glaciation by Tallman (1975)

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* Prepared in cooperation with Dr. Ann M. Tallman, Dept. of Geology, Smith College, Northampton, Mass.

and an Early Juneau/Atlin Stage by Miller, the fluctuational character is unknown, though presumably it had many glacial and intraglacial stages. Following the early Wisconsinan glaciation, the Boulder Creek Intraglacial is identified. This is based on the well-delineated peat horizon in Boulder Creek dated as "older than 40,000 years B.P. " (University of Alaska, 1973) and by the presence of old weathered till at high elevations.

After this developed the Atlin I Stage, involving thick Polar to sub-Polar ice, laying down bold lateral moraines at high level and with the lowest till overlying the Boulder Creek peat. Strong weathering on this till connotes a significant intraglacial, termed the Pine Creek Intraglacial, because of good exposures found in the gold creeks, such as Pine Creek near the 1898 gold discovery claim at Atlin. As shown in Table XV, there is a correlate of this middle Wisconsinan glaciation in the Teslin-Gladys Lake depression. This is termed the Gladys I Stage by Tallman.

Following the Atlin I/Gladys I Glaciation was the Atlin II/Gladys II Glaciation of late-middle Wisconsinan age and considered to be of sub-Polar to sub-Temperate thermophysical character on the basis of less steep lateral moraines in the terminal sector of Fourth of July Creek Valley. This glaciation is probably a correlate of the Gastineau/Sloko Stage in the Taku District. The absence of an intra-glacial in the Gastineau/Sloko Glaciation is explained by the high elevation and provenance of ice in a more maritime region where climatic conditions caused rising freezing levels and hence more likely increased glaciation in some sectors of the coastal highland.

A significant "climatic amelioration" (intraglacial ?) is indicated by a mild surface weathering of tills following the Atlin II/Gladys II Glaciation. This is believed to be the same mild weathering interface found on top of the lower till of the Gastineau Formation on the coast. Correlates of the Douglas/Inklin and Salmon Creek/Zohini Stages in the Taku District are suggested to be respectively the Atlin III Substage and the combined Atlin IV and V Substages. Correspondingly the Douglas/Inklin correlate would be the Gladys III Substage and the Salmon Creek/Zohini correlate is suggested as the Gladys IV Substage. All are considered to be late-Wisconsinan as noted in the chronology chart.

B. THE HOLOCENE

The final time interval depicted on the chronology chart is the Holocene (Table XV). Here the Thermal Maximum is equally well-documented on both the maritime and continental flanks of the Boundary Range. The Neoglacial Stage in the Taku District is two-phased, with the early phase (2500-1000 B.P.) termed the Early Mendenhall Stade (or substage ?) and the latest period of intensive mountain and cirque glaciation representing the Alaskan Little Ice Age Stade (or substage), ca. 600 years B.P. to present. The short, warm interval of the Middle Ages (900 to 1300 A.D.),so well-documented by C-14 samples from modern tills of the Mendenhall and Davidson Glaciers and from the sole of the advancing Taku Glacier, is termed the Taku Interstadial.

It is of glacio-climatological significance that the maximum morphogenetic phases of glaciation represented at the beginning of the Holocene and at the end of the Holocene (i.e., throughout the Little Ice Age) were Extended Icefield Glaciations.* Retracted Icefield Glaciations pertained during intervals of negative mass balance when low and intermediate elevation glaciers of the Stikine and Juneau Icefields took on the morphologic forms and associated terrain characteristics that we see represented by existing glaciers in these regions today (v. Figs. 5, 7 and 78).

* v. Miller (1964b) for definitions of the morphogenetic phases of glaciation.

PART XII REGIONAL CONSIDERATIONS AND THE LANDFORM DICTUM

A. TECTONIC AND CLIMATOLOGICAL ELEMENTS

This research has emphasized the role of processes in the dynamic evolution of mountain and glacial terrains. In turn this highlights the aspect of a gigantic open system with combined exogenetic and endogenetic energy inputs that are at once periodic and aperiodic. These are characterized by short-term increases in entropy of considerable magnitude and striking effect. What most distinguishes the region is its doubly critical position in a uniquely sensitive atmospheric interaction zone and in a tectonic interaction zone characterized by constant jostling of the North Pacific Oceanic plate by the Northwestern American continental plate. The result has been an intensive and continuous bombardment of cyclonic storm fronts, not equalled on earth, coupled with acutely active earthquake and mountain building stresses. The combination of high land, proximity to the sea, and turbulent maritime weather for up to 10 months each year has produced terrain of extreme relief and glaciation more extensive than anywhere outside of the Polar regions. Added to this is the unique sensitivity of the area to secular pulsations in climatic change.

These considerations pertain to the whole of the south and southeastern Alaska region and the adjoining Cordilleran areas of B.C. and the Yukon. The Juneau Icefield lies slightly east of the main zone of present day earthquake diastrophism, but is climatologically in the same regional zone. Thus, the presently glaciated area represents the fifth largest continuous icefield in North America. Because of its location near Alaska's capital at Juneau and the Yukon's capital at Whitehorse, this icefield is the most accessible for research of any in the world. This is why it has been the focus of the many previous studies documented in the references.

B. IMPLICATIONS OF GEOMORPHIC SELF-REGULATION

The coastal location coupled with the high relief of this region and the presence of high, wide-strathed valleys and plateaux, provide circumstances most propitious to the development of an icefield complex of some 32 interconnected glaciers and source névés. Because of the maritimity on the icefield's seaward flank, excessive summertime melting of ice and runoff occurs. This has led to abundant and fast-flowing streams through the coastal mountains and produced great erosion and deposition of clastics. Similarly, the wet climate and steep relief have accelerated weathering of large areas of exposed bedrock. Thus the lithologic and structural controls on topography are readily identified, as the landscape strives for dynamic equilibrium through intricate fluctuations of imposed energy...i.e., the self-regulation of its various sub-systems in the competing endogenetic and exogenetic environment.

An example of this dynamic self-regulation and what it produces is given by our studies of the processes of glacier flow through successive energy levels, as represented by (1) parabolic and (2) rectilinear (plug) and the catastrophic type known as (3) surging flow. The implication of these changes on mountain and arctic terrains as illustrating fundamental thermodynamic principles has been considered in depth by Miller (1973b) using the entropy and general systems ideas developed during the period of this ARO contract.

Longer-range examples of geomorphic self-regulation and its analysis with respect to the Juneau Icefield and adjoining regions have been given in previous studies by members of our Institute team: Miller (1963); Brew (1963); Swanston (1967) and Egan (1971). Even the orientations of cirques, berms and main glacial valleys relate directly to the orientation of regional faults and to individual joint sets in a four-fold system (Miller, 1956; Goodwin, 1973; Watkins, 1974; Seppala, 1975; Jones, 1975). To illustrate this the morphology of a prototypical hanging valley on the periphery of the Juneau Icefield is analyzed. The field work in this research was accomplished in the 1971 and 1972 summer periods of this contract.

C. THE PTARMIGAN VALLEY, A PROTOTYPE OF ROCK JOINTING AND ITS INFLUENCE ON TOPOGRAPHY *

Twenhofel and Sainsbury (1958) suggested fault control for most of the main linear valleys and ridges in this region. Miller (1963) specifically allied these to joint control, especially the steepest sets of joints. Similarly, Brew et al (1963) excluded from their analyses joints dipping less than 35° because of their small effect on orientation of major geomorphic features. Egan (1971) studied the joint patterns in the Ptarmigan valley near Camp 17 (Fig. 2) and related the topography directly to them. Included in these studies has been recognition of the strong effect of high-angle normal faulting (with associated small-magnitude graben and horst structures) on the development of topographic variations, such as gulleys and steps on the bordering ridges. Seppala (1975) has investigated the three-dimensional influence of joint sets in the Ptarmigan valley, details of which are synopsized below.

Our statistical analyses of joints have been accomplished with data obtained by readings with Brunton compasses. Foliation, which is allied to schistosity in the metamorphic rocks, has also been measured. All measurements have been plotted on stereographic nets and then transferred into contour diagrams.

The bedrock of this valley relates to Coast Range geology (Buddington, 1927; Miller, 1960). It includes marbles, lime silicates, meta-volcanics (amphibolites) and migmatitic schists with intrusive pegmatites and granodiorite zones (Forbes, 1959; Ford and Brew, 1973). The strike of schistosity in these rocks is NW, with a mean dip of 40° NE, ranging from 8° to 70°. Details of orientation of the four-set joint system at selected sites are given by Seppala (1975). In broad outline, the pattern is similar to that found elsewhere on the Icefield (Miller, 1963; Watkins, 1974), suggesting a basic relationship to regional tectonic stresses during the Cenozoic orogeny of the Alaska-Canada Coast Range. Here the two main sets strike roughly at right angles. One is northwesterly with 45-60° dips to the east; another is northeasterly with 64-90° dips to the northeast. The third set is usually vertical or even parallel to the first order set, but it dips 40° to the NE. In the analysis, the fourth set is ignored as it is essentially horizontal. The two main joint sets form a well-developed cross joint network which facilitates and controls the movement and impounding of free water and hence of frost weathering.

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* This section prepared in cooperation with Dr. Matti Seppala, Department of Geography, University of Turku, Finland, of our 1971 field team.

Micro-relief Forms and the Valley's Assymetry

In cross-section the Ptarmigan valley is assymetric, with a steep west-facing slope (70°) and gentler east-facing slope (35°). A similar configuration is found in the upper (glacier filled) Lemon Creek Glacier valley (Figs. 8a and 75).

Micro-relief features show the effect of small faults and differing lithologies, with weathered out blocks reflecting all four joint [illegible] In such intermediate and high elevations in this intensely ma[illegible] dominant weathering is chemical breakdown, strongly abetted by [illegible] shat[illegible] above 3000 feet the environment is periglacial. In the Miocene and Pli[illegible] much pre-glacial fluvial erosion and down-dip mass wastage of weathered ma[illegible] helped to initiate the assymetry of the valley. Pleistocene and Holocene glaciers have overdeepened the valley and periodically removed material mass-wasted from the valley walls. But, it is the bedrock structure that has exerted the primary control.

It is significant that the strike of the first order joint set almost parallels the axial direction of the valley. As foliation also parallels the predominant set, this has provided a strong control on the valley's orientation. The steepness (nearly 90°) of the second and third order joints has been the main element influencing the shape of the valley.

Formation of the Ptarmigan valley and of the adjacent Lemon Glacier and Canyon Creek valleys to the east and west has been assymetric because the predominant joint set dips at an angle which is diagonal to the flow direction of the glacier (Fig. 75).

A Conceptual Model

Seppala has examined the formation of such valleys theoretically, with simple models of different possible joint sets and of their to be expected influence on valley form. For example, if the surface of a predominant joint set is found to conform to one slope of a valley, where a second and weaker set dips 90° to the predominant one it is logical that valley assymetry will develop whenever the dip of the first-order set lies close to 0°, 45° or 90° (see A, B and C in Fig. 76). Such geometries produce optimum conditions for forming a regular (symmetric) U-form glacial valley. In other cases (see D and E, Fig. 76), a valley should have a more or less assymetric cross-section. If the joints are not perpendicular, of course, a more complex morphology will result.

From the flow law of ice it can be demonstrated that the paraboilc characteristic of most glacier flow curves (in low and intermediate entropy ranges) when turned on their sides do reveal U-form cross-sections. As this represents the true nature of the energy distribution, it will lead to U-form valley profiles. Because ice is a medium which is visco-plastic (if thermophysically Temperate) or elasto-plastic (if thermally Polar), it follows the topography in such a way that maximum erosive power is directed to the bottom of the glacial valley. There it will capitalize on any zones of structural or lithologic weakness. As the Ptarmigan Glacier transects a variety of rock types of variable resistances, it follows that dominant control on the development of the valley form has been bedrock structure. The influence of lithology is

mostly in the micro-relief terrain features of the upper flanks and ridge crests between the glacial valleys. These concepts are verified in most other valley and ridge areas of the Juneau Icefield and its peripheral Taku and Atlin Districts (e.g., Fig. 78).

D. REGIONAL DISEQUILIBRIUM AND THE MOST PROBABLE STATE

In the foregoing, the principle of dynamic equilibrium in landform and terrain analysis applies. But the effects of pre-existing controls (structural and lithologic) end up dominating the terrain evolution. As such controls impose constraints...i.e., lend themselves to asymmetry in the Ptarmigan Glacier valley...that particular landform can only be looked upon as striving for the most probable state (Leopold and Langbein, 1962). In a few other valleys in this region where there is more lithologic homogeneity and where similarly orientated joint sets exist on either valley wall, more symmetrical U-form valley shapes occur. As in the ultimate symmetry produced by the Law of Divides, such cases have reached dynamic equilibrium and hence represent a steady state. In this region of such varied rock types and vast differences in relief, such is seldom seen.

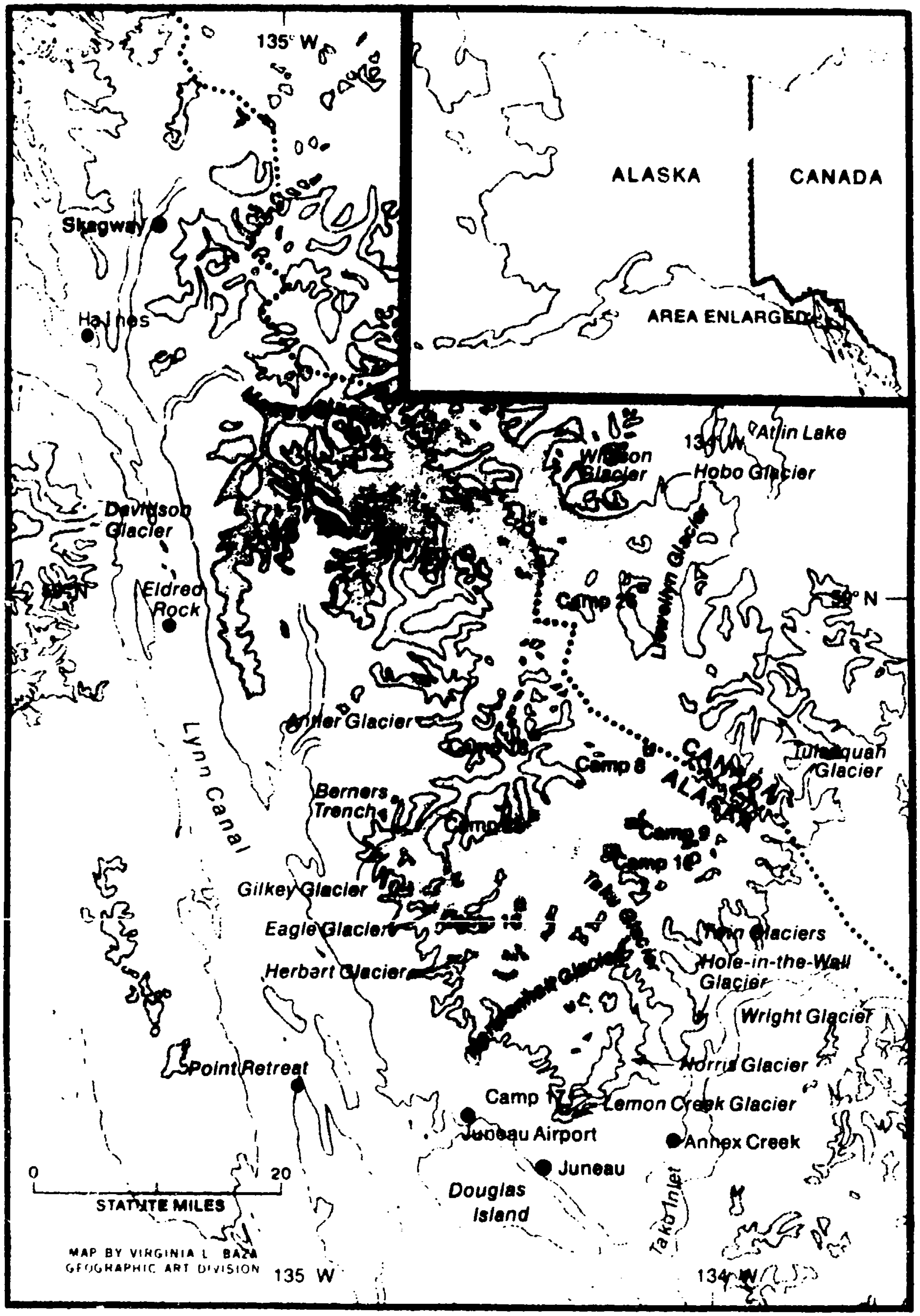
Considering the variety and magnitude of relief on the icefield (Figs. 5,7,78) and the relatively short time-span which encompasses the Pleistocene (ca. 3.5×10^6 years), we can only conclude that the intensity of the glacial, glacio-fluvial, mass wastage and periglacial processes which have so much been the focus of our attention in this report, has been so great that the entropic evolution of the terrain features described has been exceedingly rapid. This has only been achieved with a considerable amount of crustal and atmospheric instability. Thus the rapidity of erosion has been directly a consequence of the unique combination of exogenetic and endogenetic factors described. As long as these kinds and rates of energy input into the terrain system remain high, this northwestern coastal region will remain in dynamic disequilibrium, quite in contrast to the dynamic equilibrium of low relief regions in the mid-continent sectors of North America.

Table XV. Provisional Chart of Wisconsinan and Holocene Stratigraphic Sequences in the Northern Boundary Range, Alaska – Canada with Suggested Comparisons with Other Regions.

YRS BP x 1000	COOK INLET AFTER KARLSTROM, 1964	ALASKA COMPOSITE PÉWÉ 1975	SOUTH-WESTERN YUKON TERRITORY AFTER DENTON AND STUIVER, 1967	EASTERN YUKON TERRITORY AFTER BOSTOCK 1963, HUGHES ET AL., 1969	RELATIVE EXTENT AND CHARACTER ATLIN REGION GLACIATIONS	FOURTH OF JULY CREEK VALLEY, ATLIN REGION TALLMAN, 1975	TAKU DISTRICT S.E. ALASKA AND N.W. BRITISH COLUMBIA MILLER, 1956, 1969 AND 1975
0–2		IV	Neoglaciation	Neoglaciation	Re-occupation of Nivation Hollows and Cirques ?	Holocene: High Level Permafrost; Nivation Hollows; Palsa Development	Holocene – Neoglacial: Alaska Little Ice Age; Taku Intrastadial; Early Mendenhall
2–8		III	Slims Nonglacial			Thermal Maximum	Thermal Maximum
8–10	Naptowne Glaciations: Tanya				Glacial Fluvial Terraces	Intense Periglacial; Atlin V, Atlin IV / Gladys III	Salmon Cr / Zoh ...: High Cirque Glaciation*; Sittakanay Substage**; Tulsequah Substage***; King Salmon Substage****
10–13	Skilak Lake				Temperate to Sub-temperate; Weathered Tills ?	Atlin III / Gladys III; Significant Climatic Amelioration	Douglas/Inklin Stage; Intraglacial
13–20	Killey	II	Kluane Glaciation	McConnell Glaciation	Sub-temperate to Sub-polar ?; Strong Weathering Ontills ?	Atlin II / Gladys II	Greater Mountain Ice sheet: Gastineau/Sloko Stage
20–28	Mooshorn				Thick Polar to Sub-polar ?	Gladys I; Pine Creek Intraglacial; Atlin I	(Glacial and Intraglacial Substages)
28–45	Knik		Boutellier Nonglacial; Icefield	Reid Glaciation	Old Weathered Drift Overlain by Till ?	Boulder Creek Intraglacial	Intraglacial (relatively cool)
45–80+		I	Silver Nonglacial; Shakwak		Character Unknown Probably Many Glacials and Intraglacials ?	Pre-Atlin I (Pre-Classical Wisconsinan)	Greater Mountain Ice sheet: Early Juneau/Atlin Stage (Probably extends into classical upper Sangamonian)

*Retracted Icefield Glaciation
**Extended Icefield Glaciation
***Lesser Mountain Ice sheet
****Intermediate Mountain Ice sheet

YRS BP x 1000	NORTH-CENTRAL AND EASTERN BRITISH COLUMBIA AFTER RUTTER, 1975	COLORADO FRONT RANGE AFTER MADOLE, MAHANEY, AND FAHEY, 1975	PUGET LOWLAND AFTER EASTERBROOK, 1963 AND 1975, AND CRANDELL, 1965	GREAT LAKES COMPOSITE AFTER FLINT, 1971, AND TERASMAE AND DREIMANIS, 1975	WISCONSIN AND ILLINOIS AFTER BLACK, 1975, AND FRYE AND WILLMAN, 1973
0–2	(No dates or correlations given)	Gannet Peak; Audubon	Alpine Glaciation; Alpine Glaciation		
2–8		Triple Lakes			
8–10		Pinedale: Late Stade	Sumas Stade (Washington State)	Valderan	Valderan Glacial
10–14	Deserts Canyon	Middle Stade	Everson Interstade; Vashon Stade	Two Creekan; Mankato; Cary	Two Creekan Intraglacial
14–22	Lake Portage Canyon; Early Portage Mountain	Early Stade	Interstadial; Evans Creek Stade	Tazewell; Iowan	Woodfordian Glacial Stage (3 main till units, with retreatal deposits)
22–28			Olympia Interglacial	Farmdalian Intraglacial	Farmdalian Intraglacial
28–45	Early Advance possibly pre-Wisconsinan	Bull Lake - Pinedale Interglacial ? ? ?	Salmon Springs II Stade; Salmon Springs Interglacial	Upper Altonian	Altonian Glacial Stage (Five main tills, with retreatal deposits)
45–75		Bull Lake (Part of Sangamonian ?)	Salmon Springs I Stade	Port Talbot Intraglacial; Lower Altonian	
75+			Puyallup Interglacial Sangamonian?	Sangamonian	Sangamonian



JUNEAU ICEFIELD RESEARCH PROGRAM GLACIOLOGICAL STATIONS

U. S. WEATHER BUREAU METEOROLOGICAL STATIONS

FIG. I LOCATION MAP OF THE JUNEAU ICEFIELD IN SOUTHEASTERN ALASKA

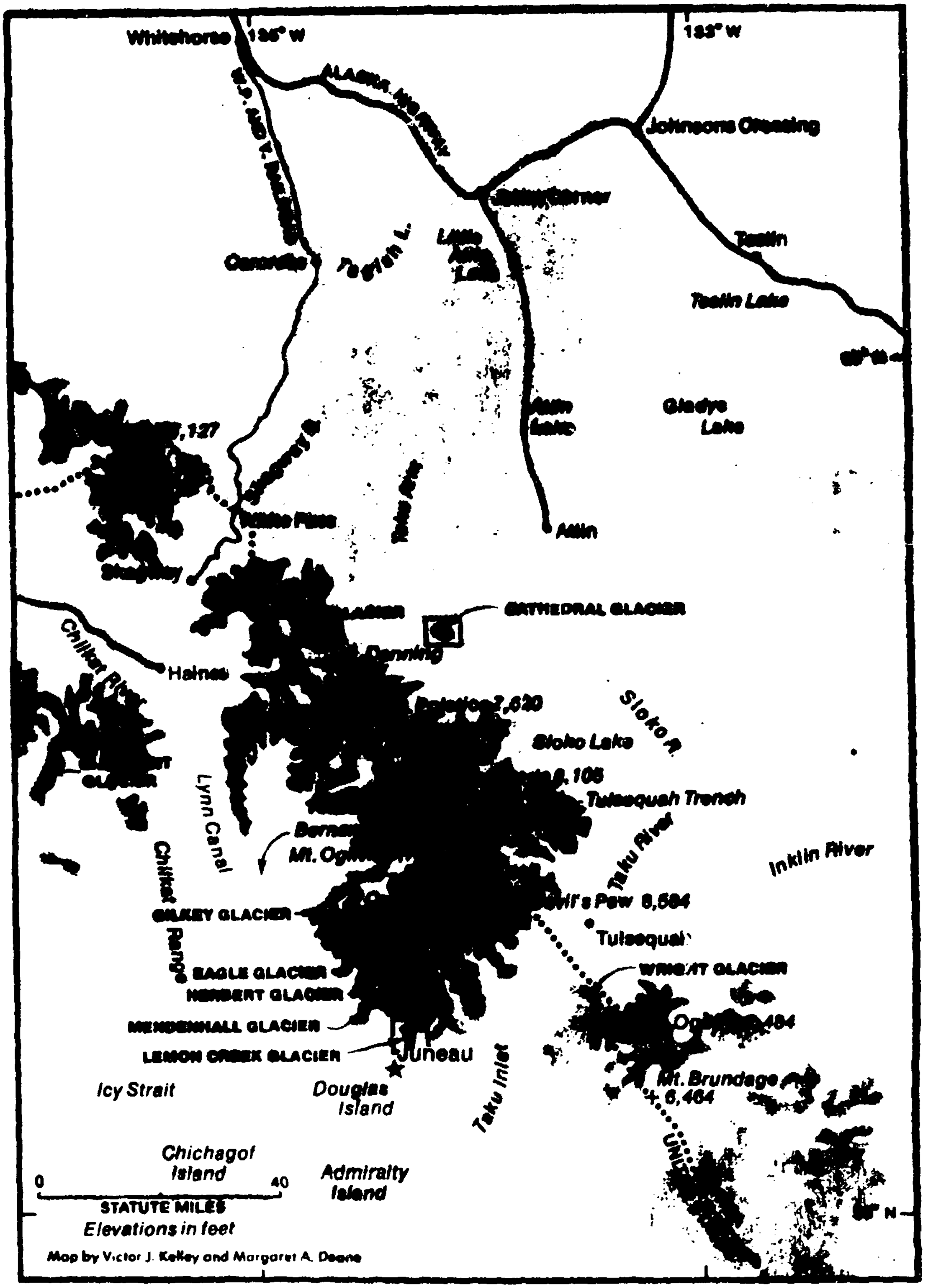
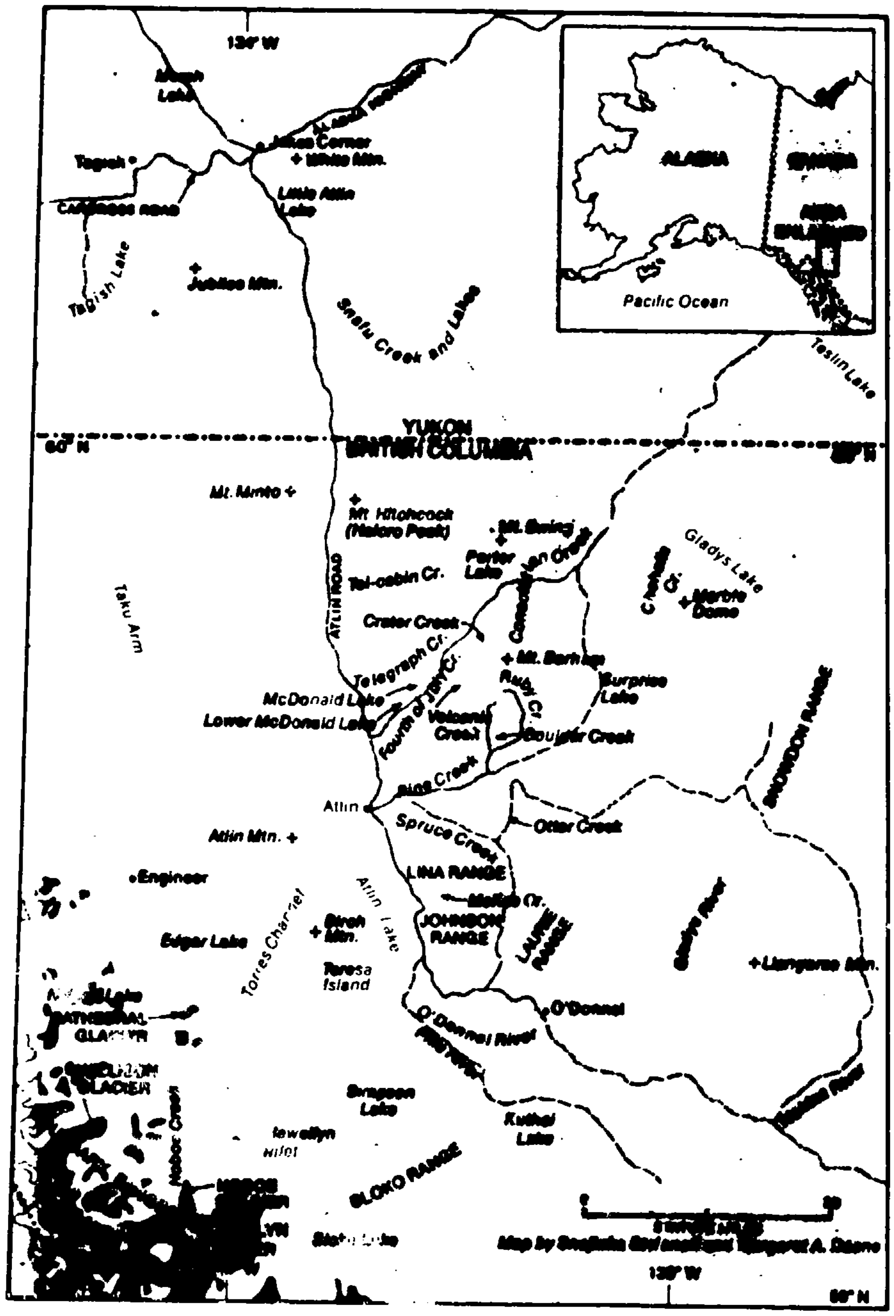


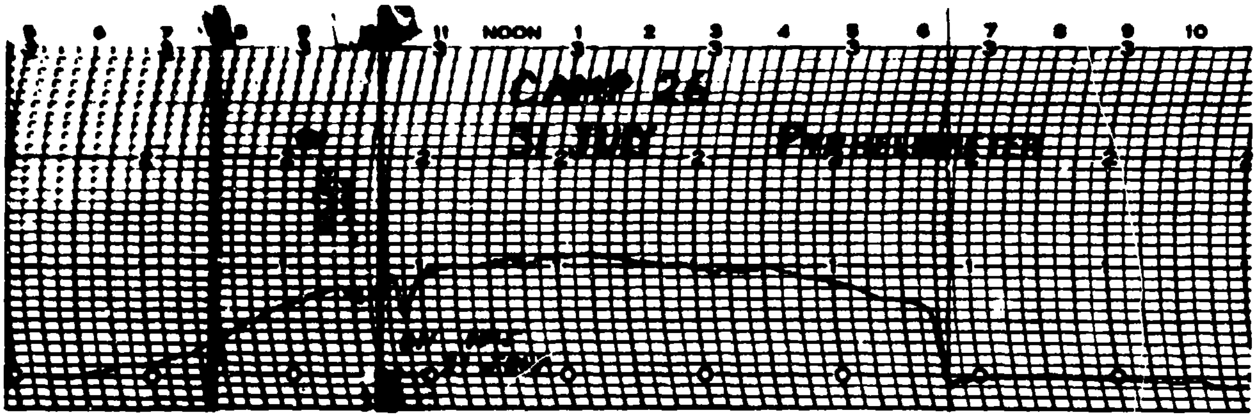
FIG.2 NORTHERN BOUNDARY RANGE AND JUNEAU ICEFIELD IN TAKU AND ATLIN DISTRICTS



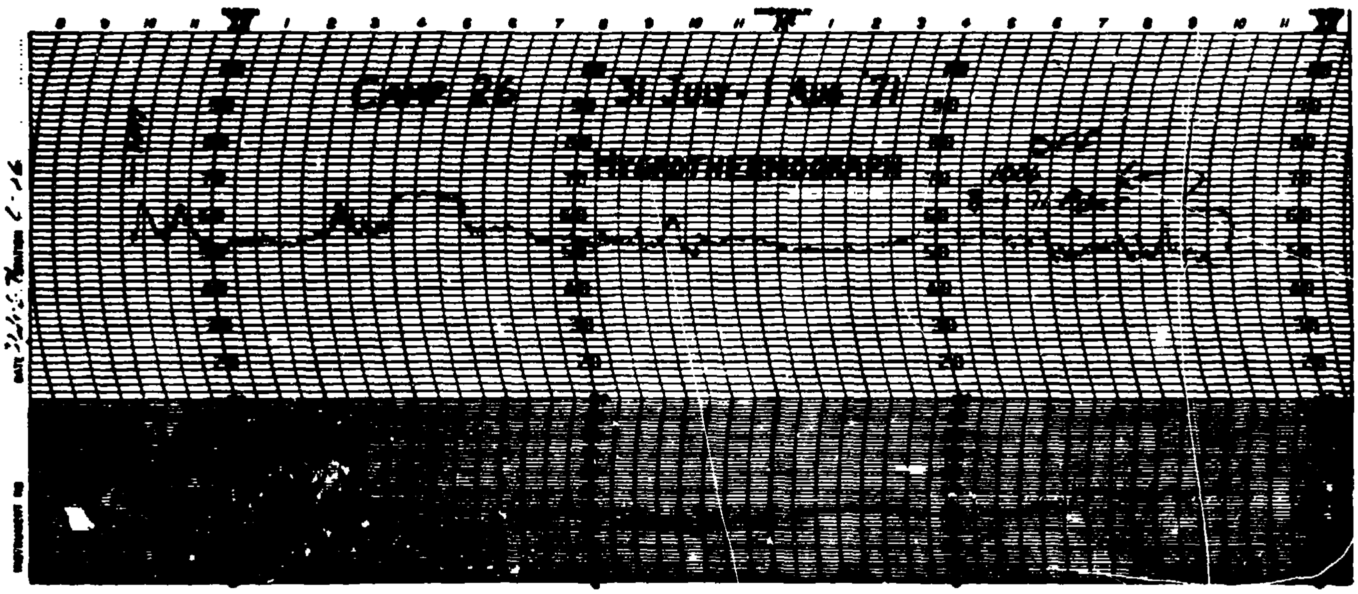
Fi. 3 Map of the Atlin Region (Courtesy of the National Geographic Society)

Fig. 4 Satellite visual from NOAA-4 on 23 April 1975 at 12:53 p.m. Altitude 1500 Km. View shows southeast Alaska Coast on right with storm front associated with North Pacific low in center covering southwest Alaska and the Bering Sea. Note locations of Anchorage, Yakutat and Juneau.

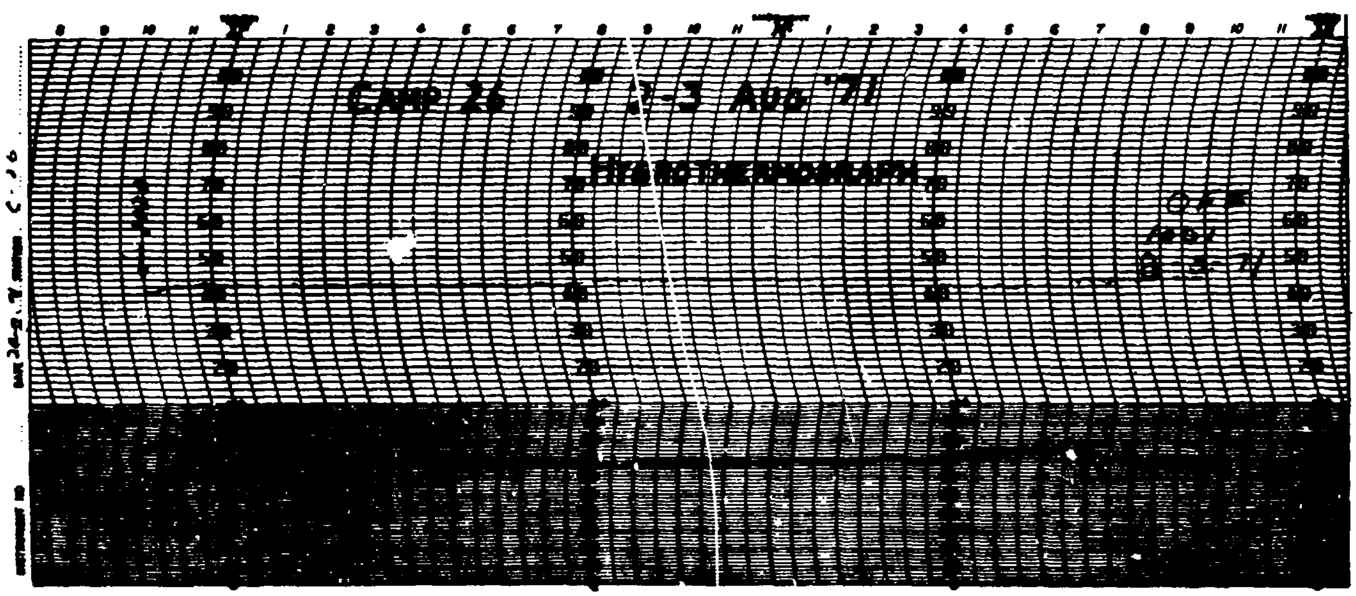
Fig. 5 Oblique aerial view looking southwest across southern portion of the Juneau Icefield



a. Pyrheliograph record for 31 July 1971.



b. Hygrothermograph trace for 31 July 1971.



c. Hygrothermograph trace for 2 August 1971.

Fig. 6 Selected Meteorclogical Charts, Camp 26.

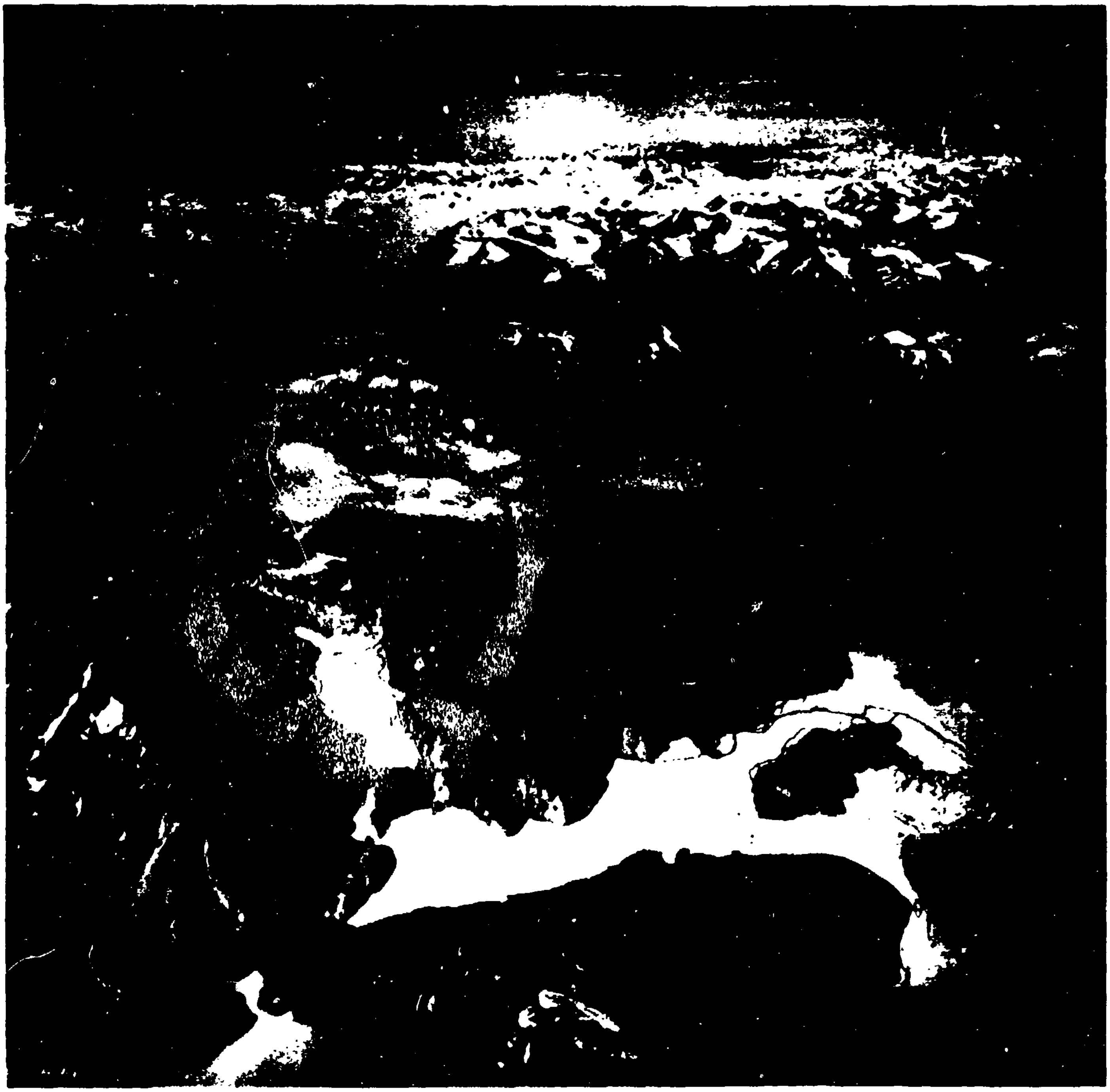


Fig. 7

View west across terminus of Llewellyn Glacier, between Camp 26 and Llewellyn Inlet showing part of icefield periphery under study in Atlin sector. (B. C. Government photo, August 30, 1949)

a. Lemon & Ptarmigan Glaciers near Juneau -- view southeast on August 13, 1948. Part of Thomas Glacier in left foreground (photo M. M. Miller)

b. Coliseum and Cathedral Glaciers near Atlin – view southwest on July 14, 1975. Torres Rock Glacier in hanging valley on left (photo M. M. Miller)

Fig. 8 Prototype coastal and inland cirque - headed Glacier systems showing research camp locations.

Fig. 9 Cathedral Massif showing Chapel and Cathedral glaciers in center (left and right). Torres Rock Glacier Valley, lower right; Cathedral Creek, upper right. North at top (photo courtesy Canadian Department of Mines, Energy and Resources).

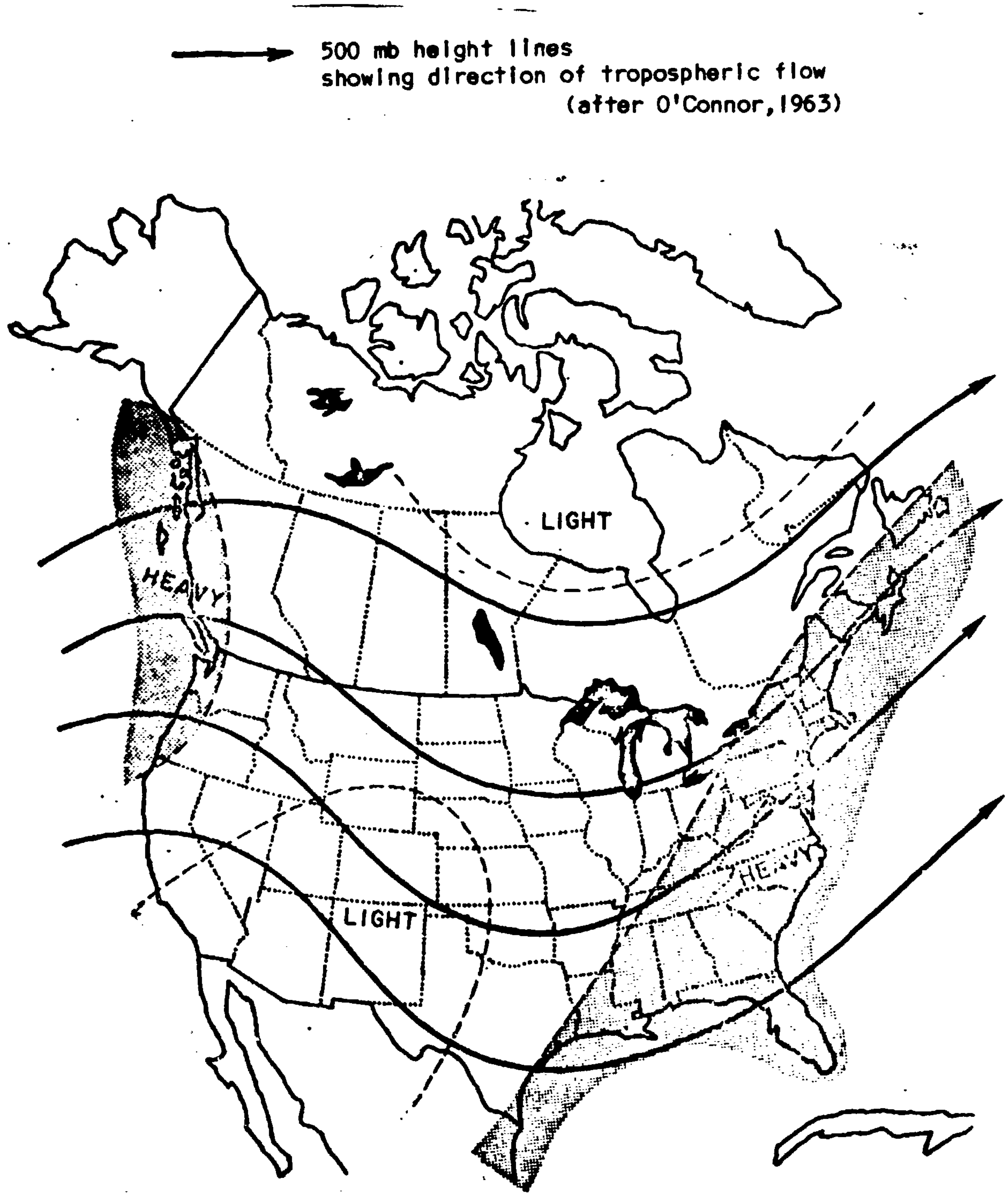


Fig. 10 Relationship of Upper Troposphere Pressure Patterns to Precipitation

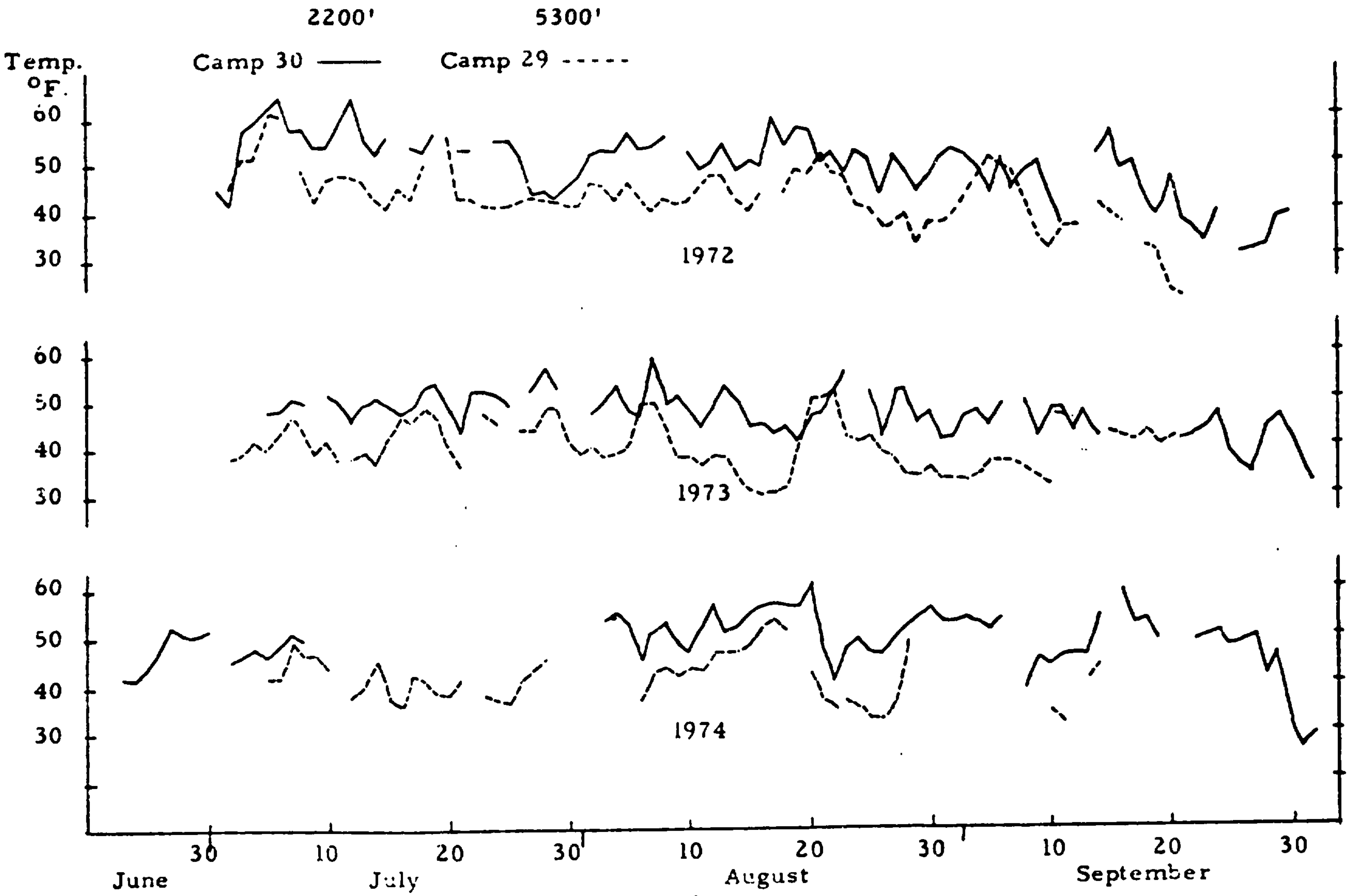
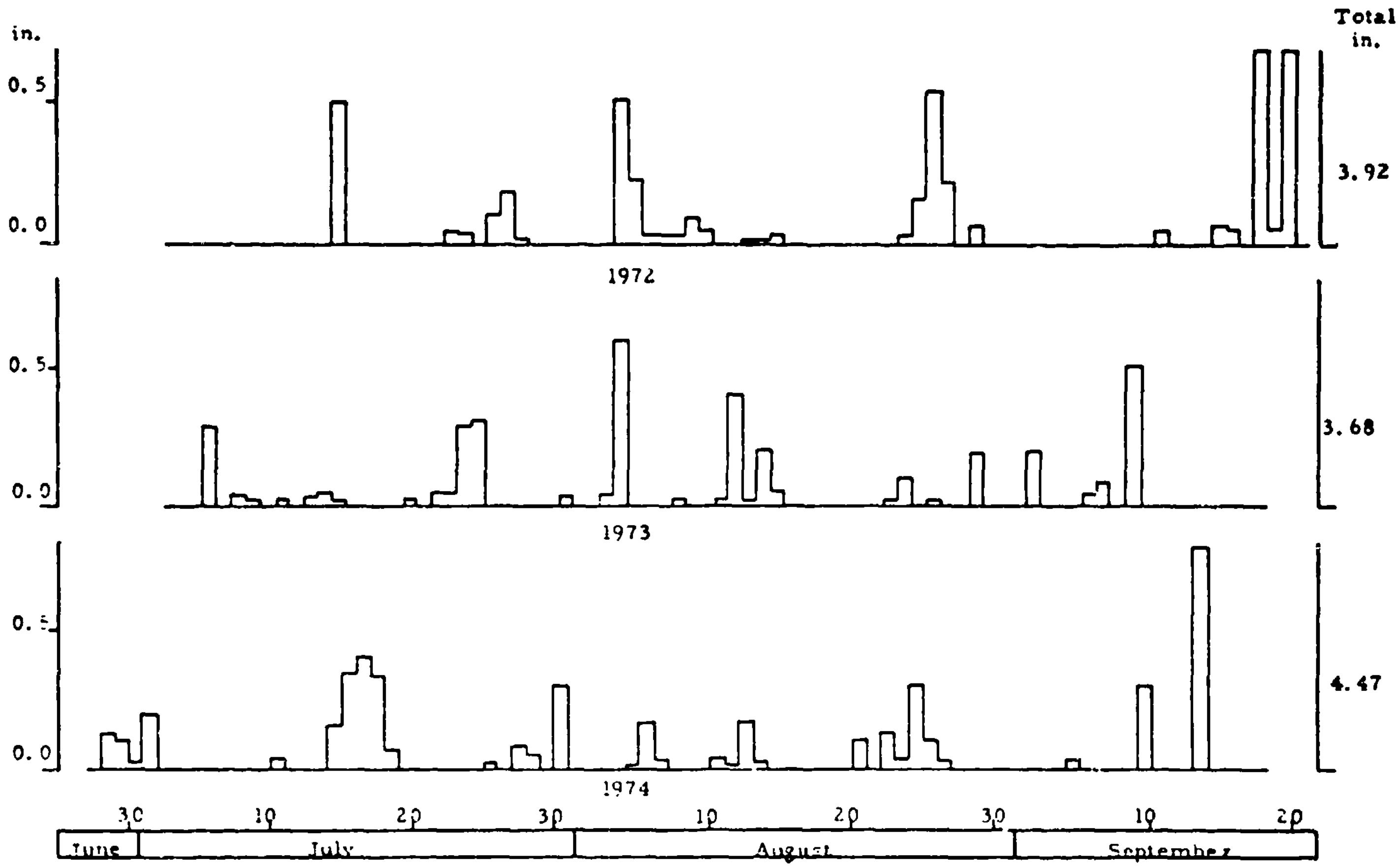
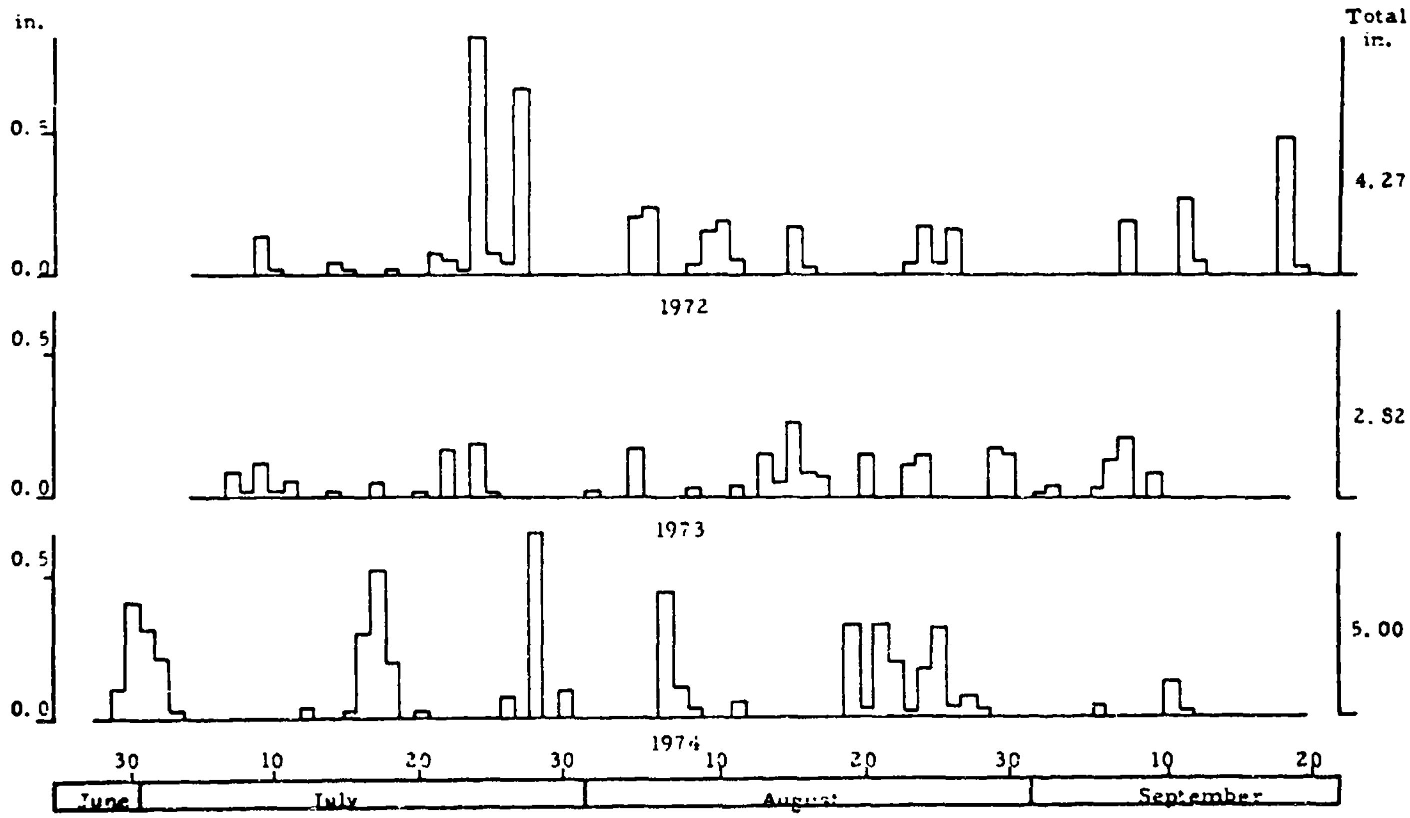


Fig. 11 Daily Temperatures, Camps 29 and 30, 1972 - 74 Summer Field Seasons



a. Daily Precipitation, Camp 29 (Cathedral Station), 1972-1974 Summer Field Seasons



b. Daily Precipitation, Camp 30 (Atlin Station), 1972-1974 Summer Field Seasons

FIG 12. COMPARISON OF SUMMER PRECIPITATION PATTERNS FOR CAMPS 29 & 30 IN THE ATLIN SECTOR

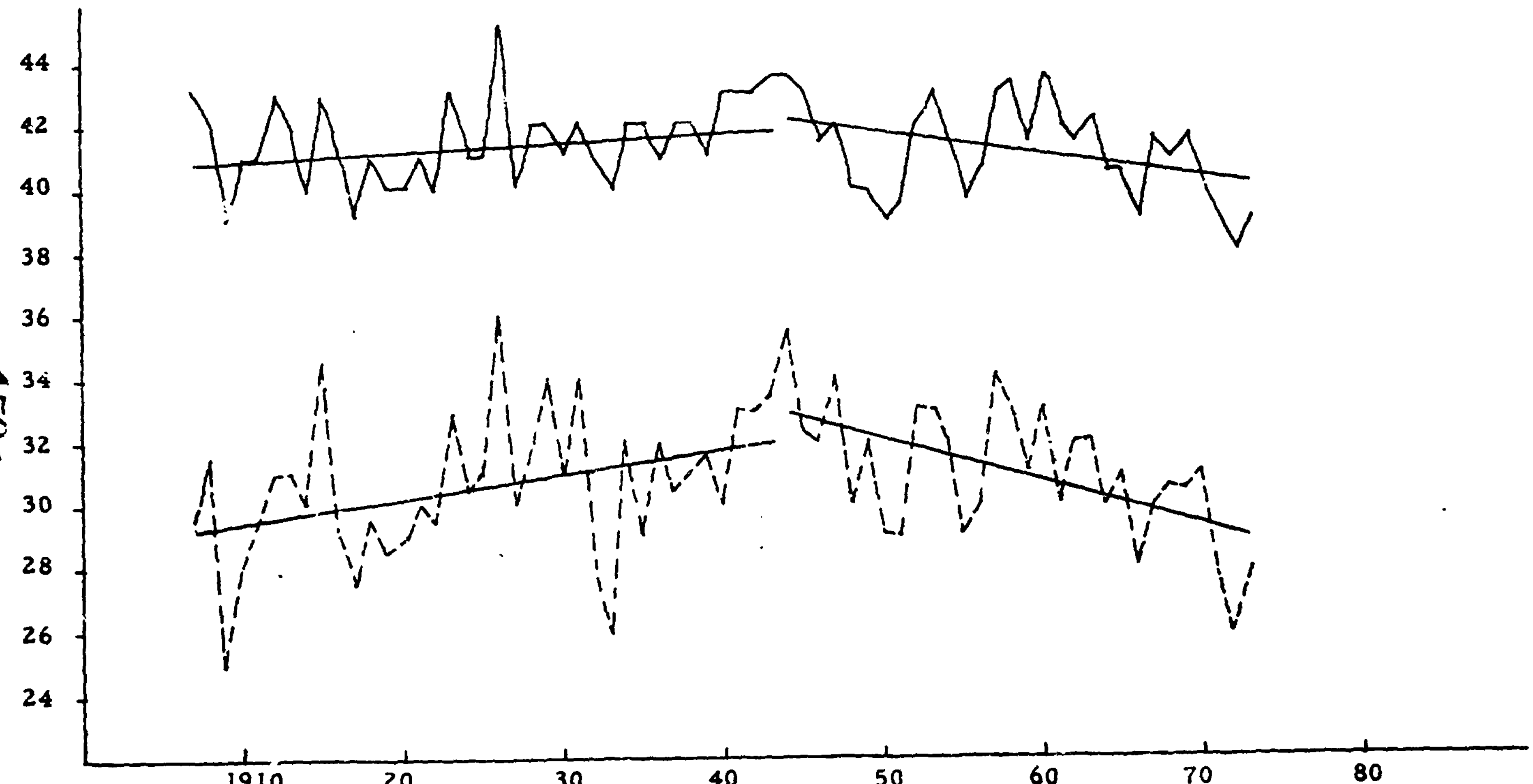


Fig.13 Annual Mean Temperatures, Coast and Interior, 1907-1973, With Linear Regressions y=a+bx

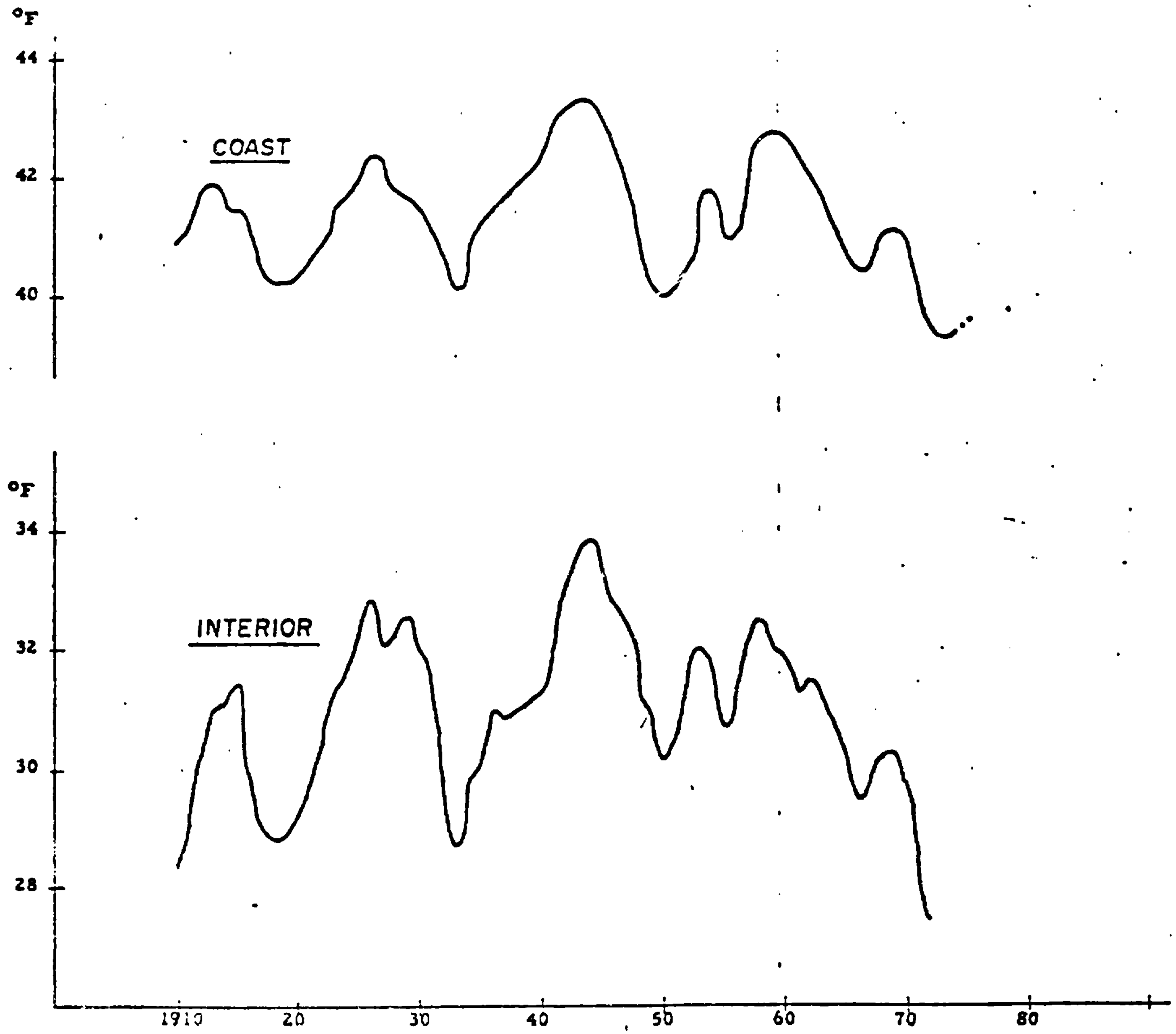


Fig. 14 Temperature, 5-year Weighted Annual Means, Coast and Interior, 1907-1973

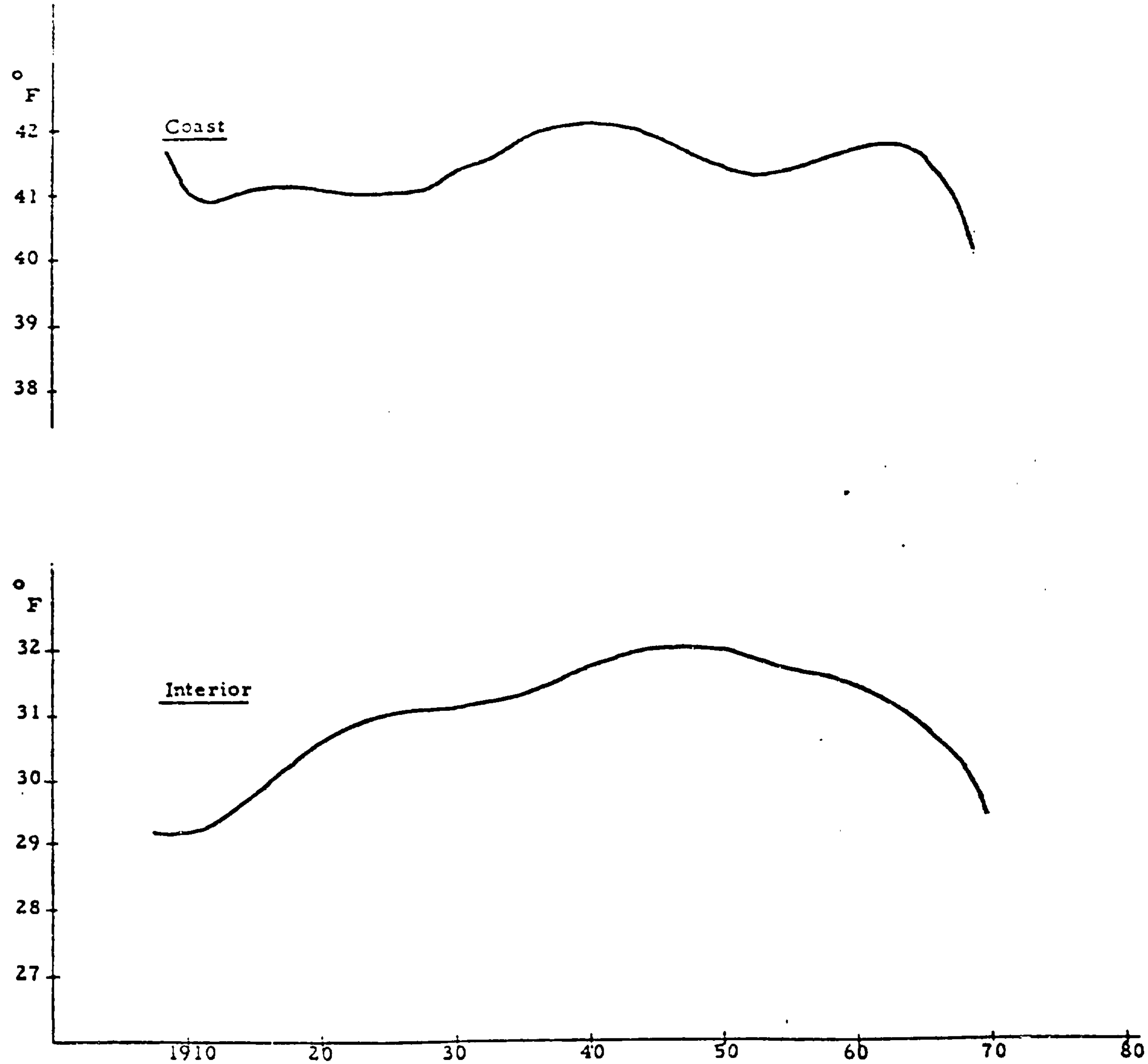


Fig.15 Polynomial Regression of Temperatures, 1907-1973

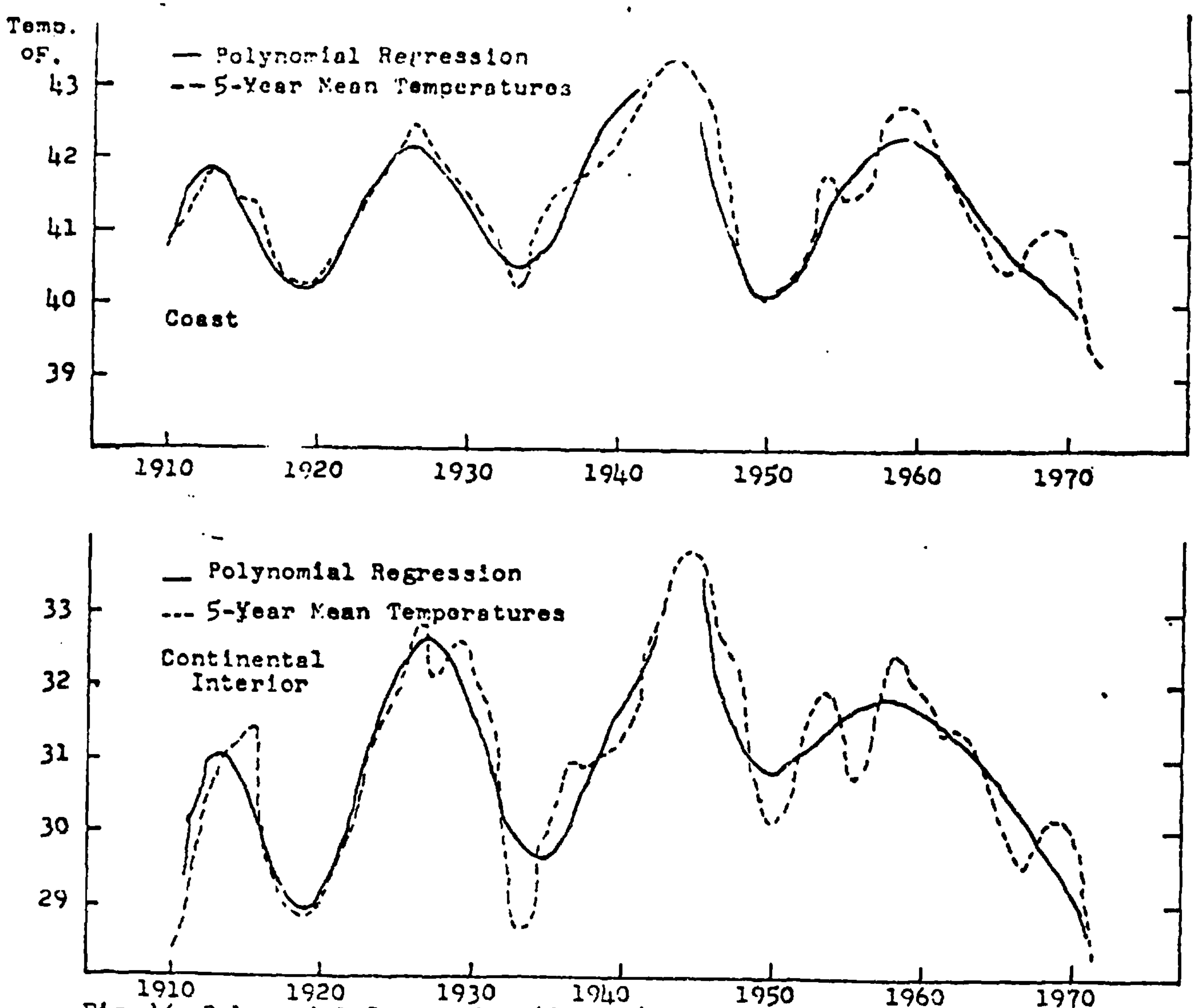


Fig. 16 Polynomial Regression (2-part) and Five-Year Weighted Moving Average of Mean Annual Temperatures, Coast and Continental Interior

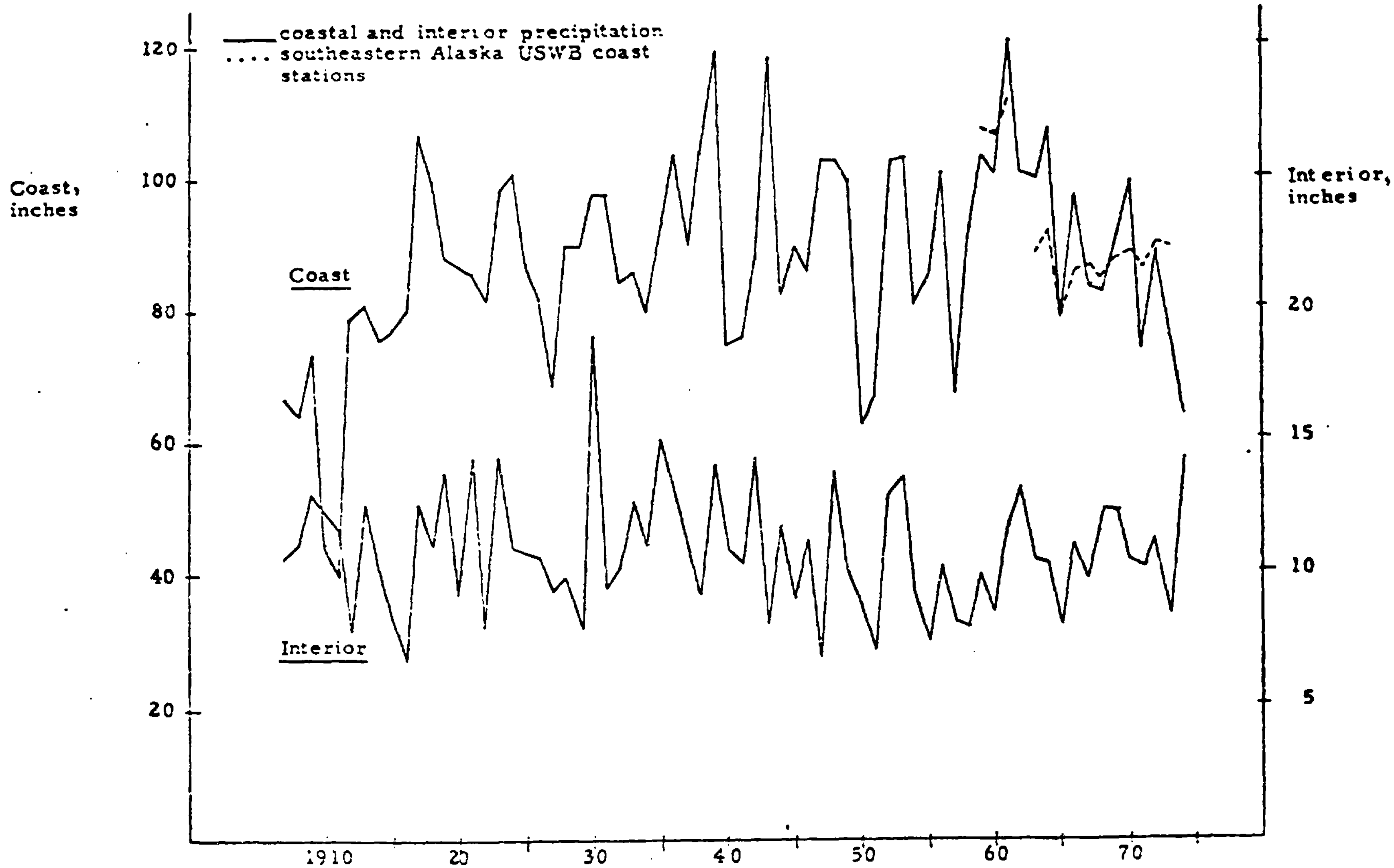


FIG. 17 COMPARISON OF ANNUAL PRECIPITATION FOR COAST AND INTERIOR AREAS, 1907-1973

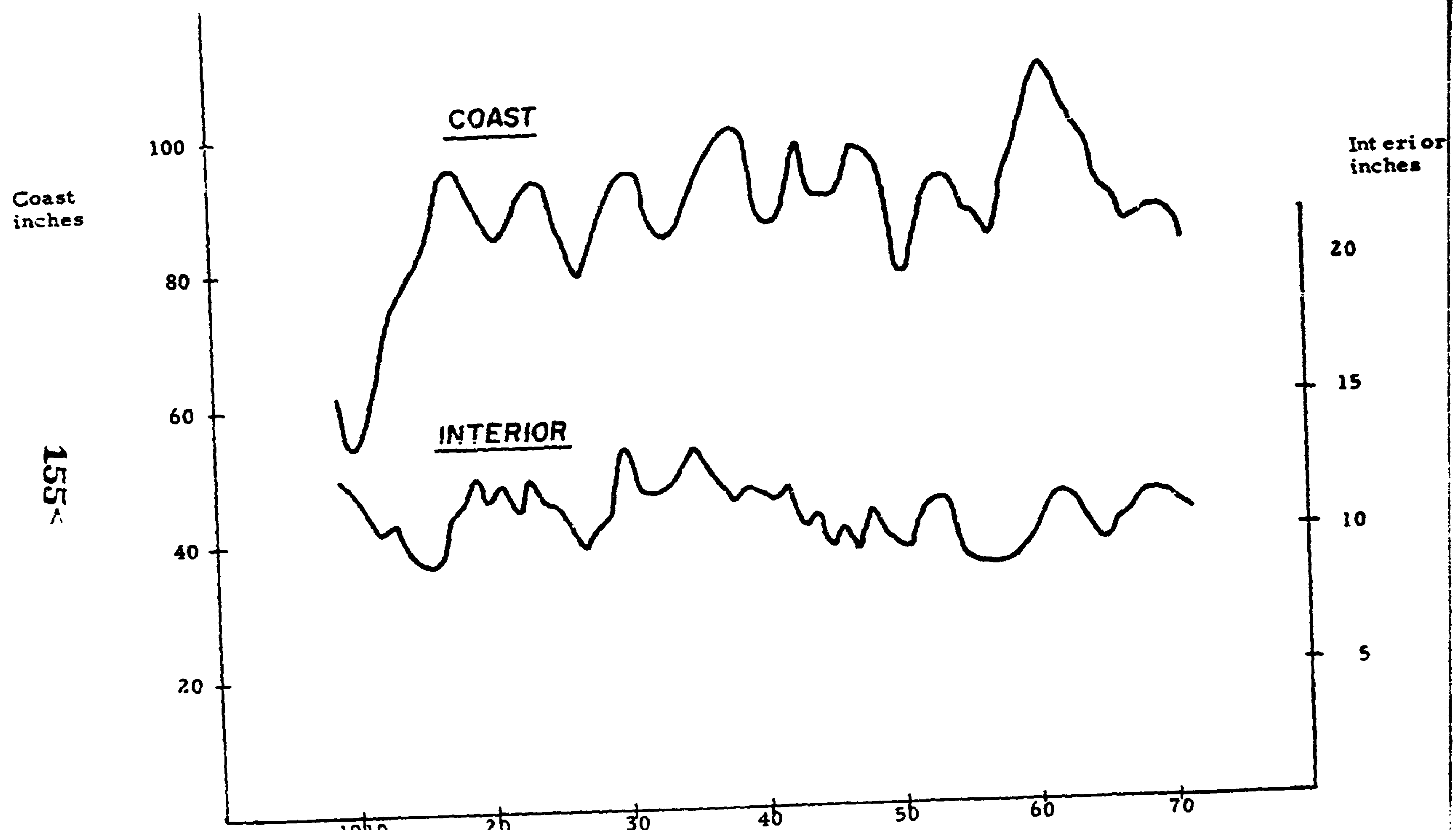


Fig. 18 Annual Precipitation, 5-year Weighted Means, Coast and Interior, 1907-1973

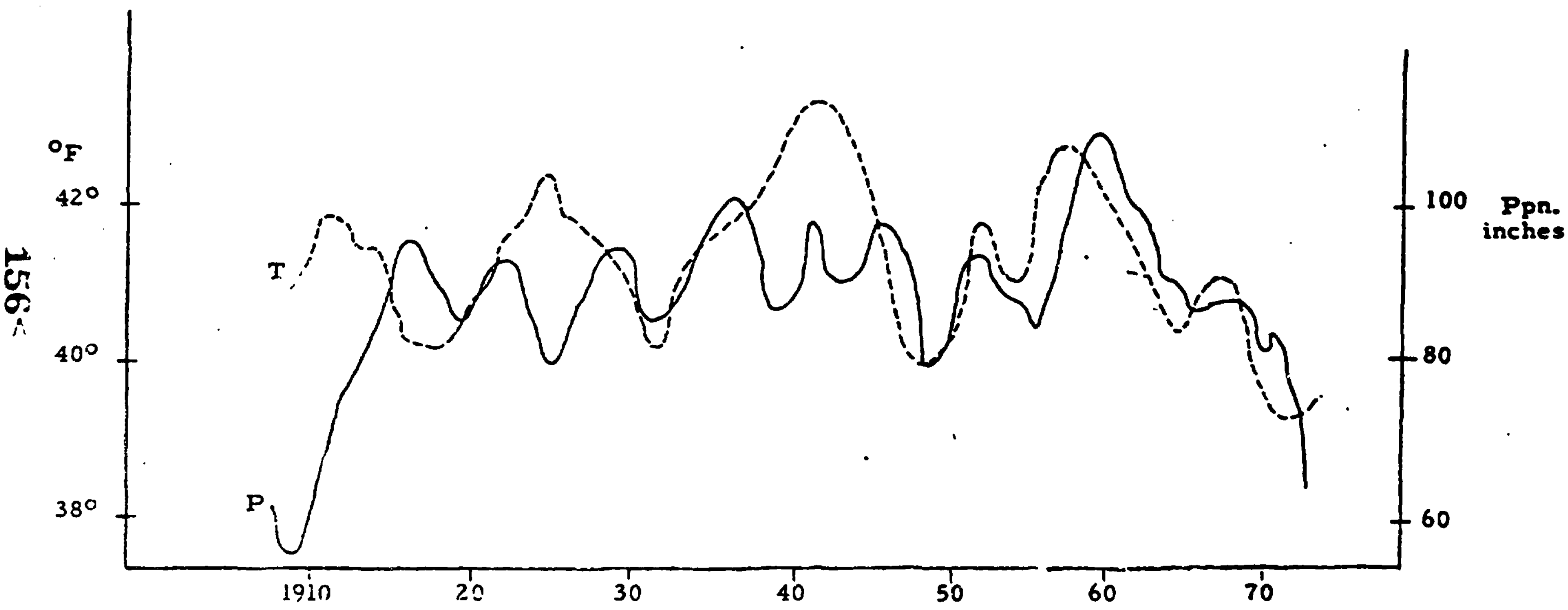


Fig. 19 Annual Temperature and Precipitation, Coast, Five-Year Weighted Means, 1907 - 1973

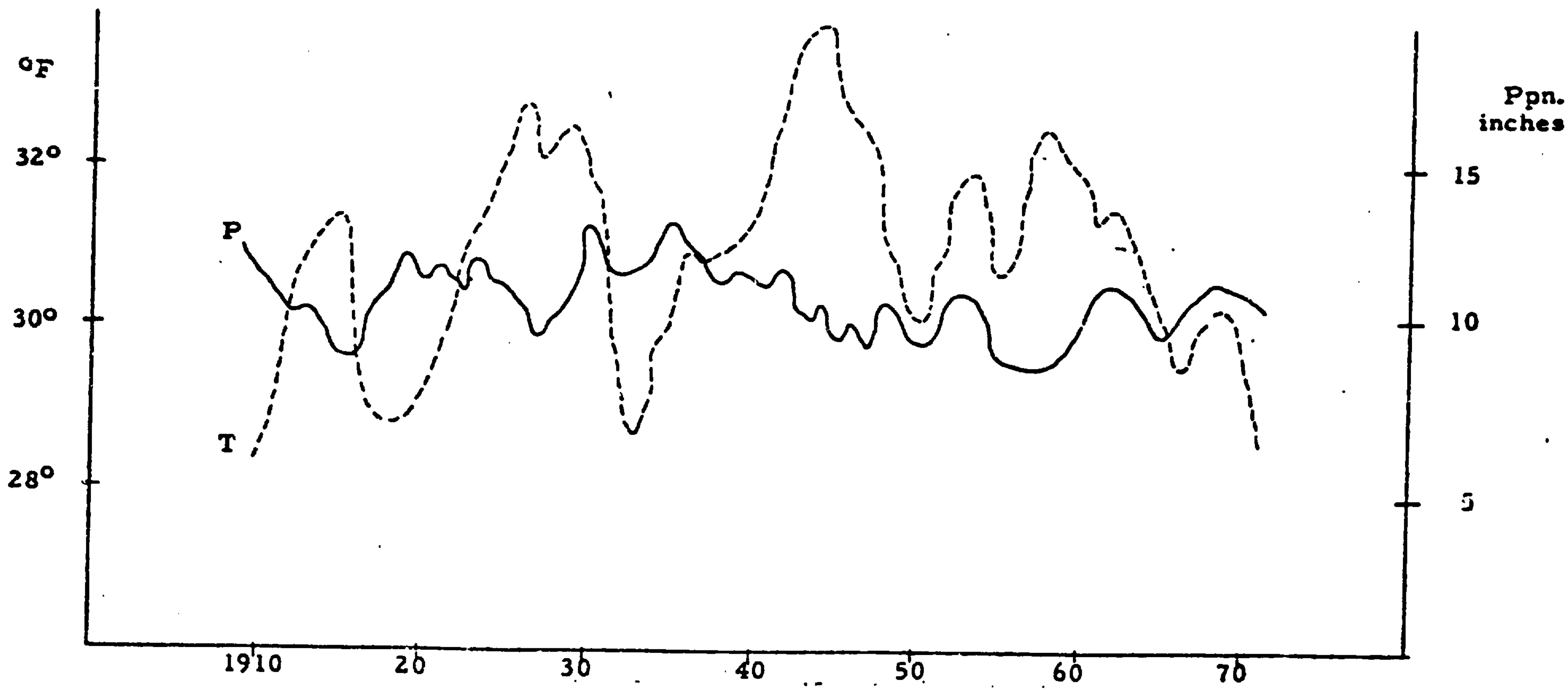


Fig. 20 Annual Temperature and Precipitation, Interior, Five-Year Weighted Means, 1907 - 1973

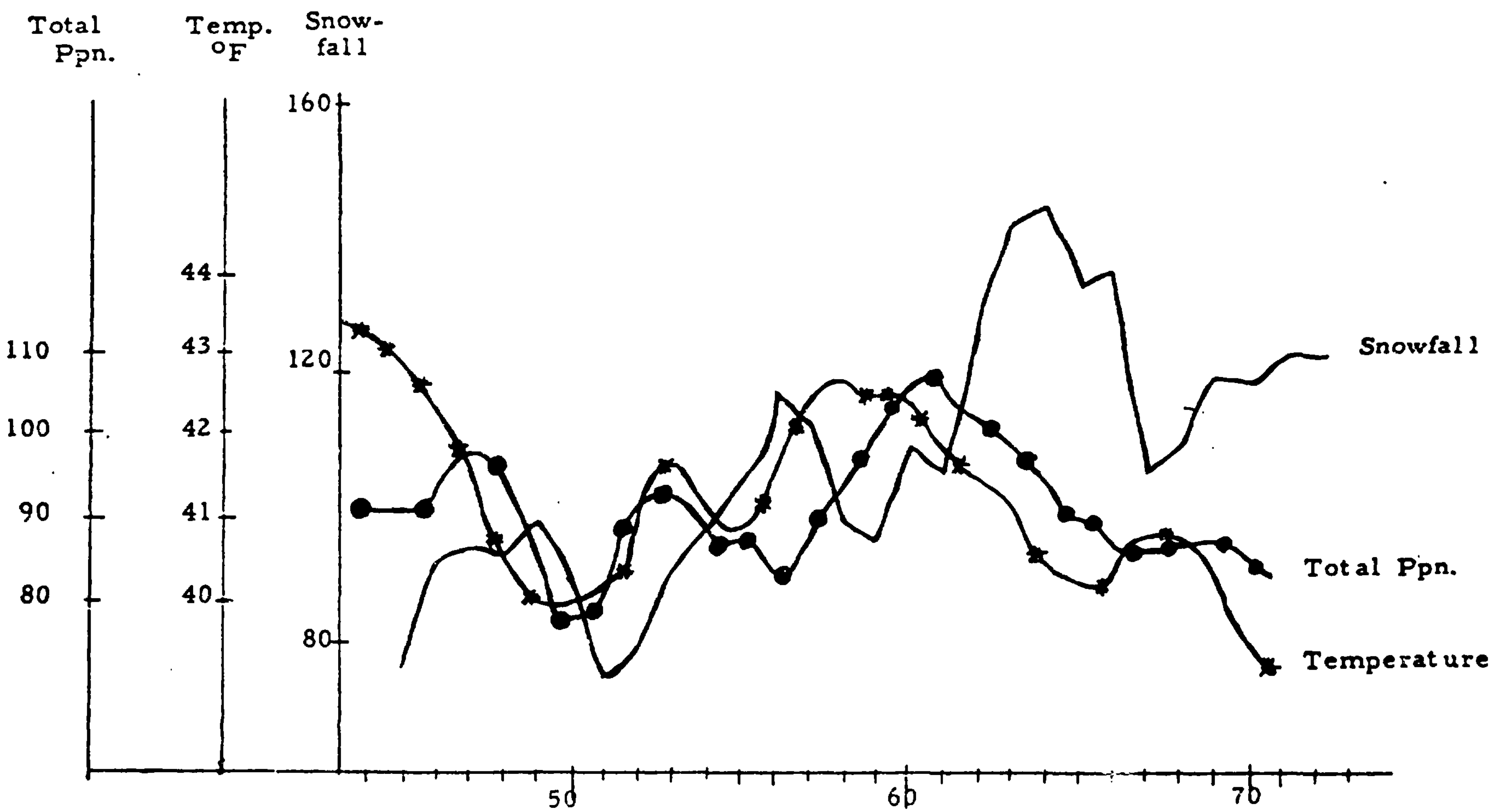
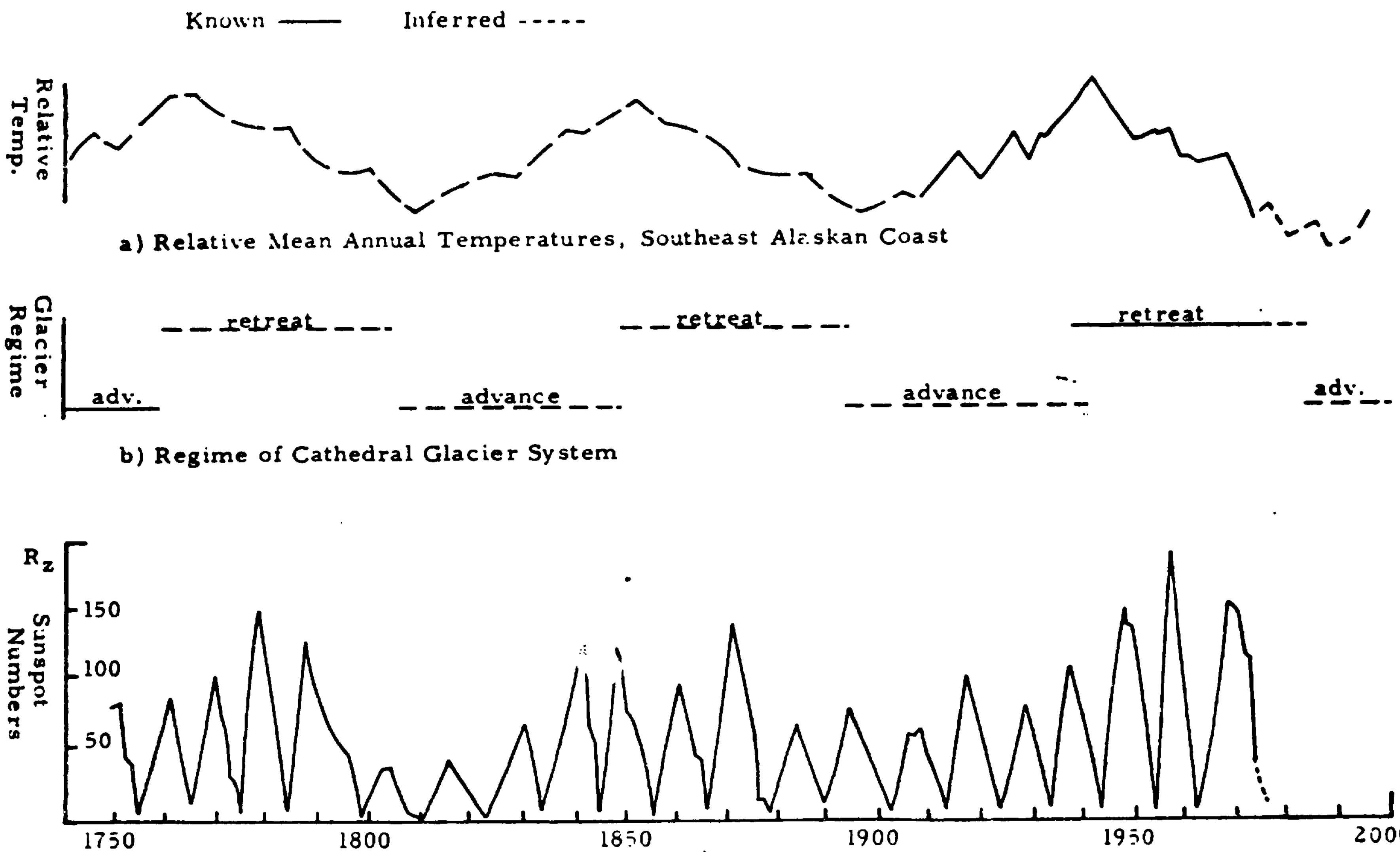


Fig. 2) Five-year Means of Total Snowfall, Juneau Airport, and Total Precipitation and Temperature, Coastal Region, 1943-1974



a) Relative Mean Annual Temperatures, Southeast Alaskan Coast

b) Regime of Cathedral Glacier System

c) Annual Relative Sunspot Numbers (R_z)

Fig. 22 Sunspots, Temperatures, and Cathedral Glacier Advance and Retreat, 1750 - 2000

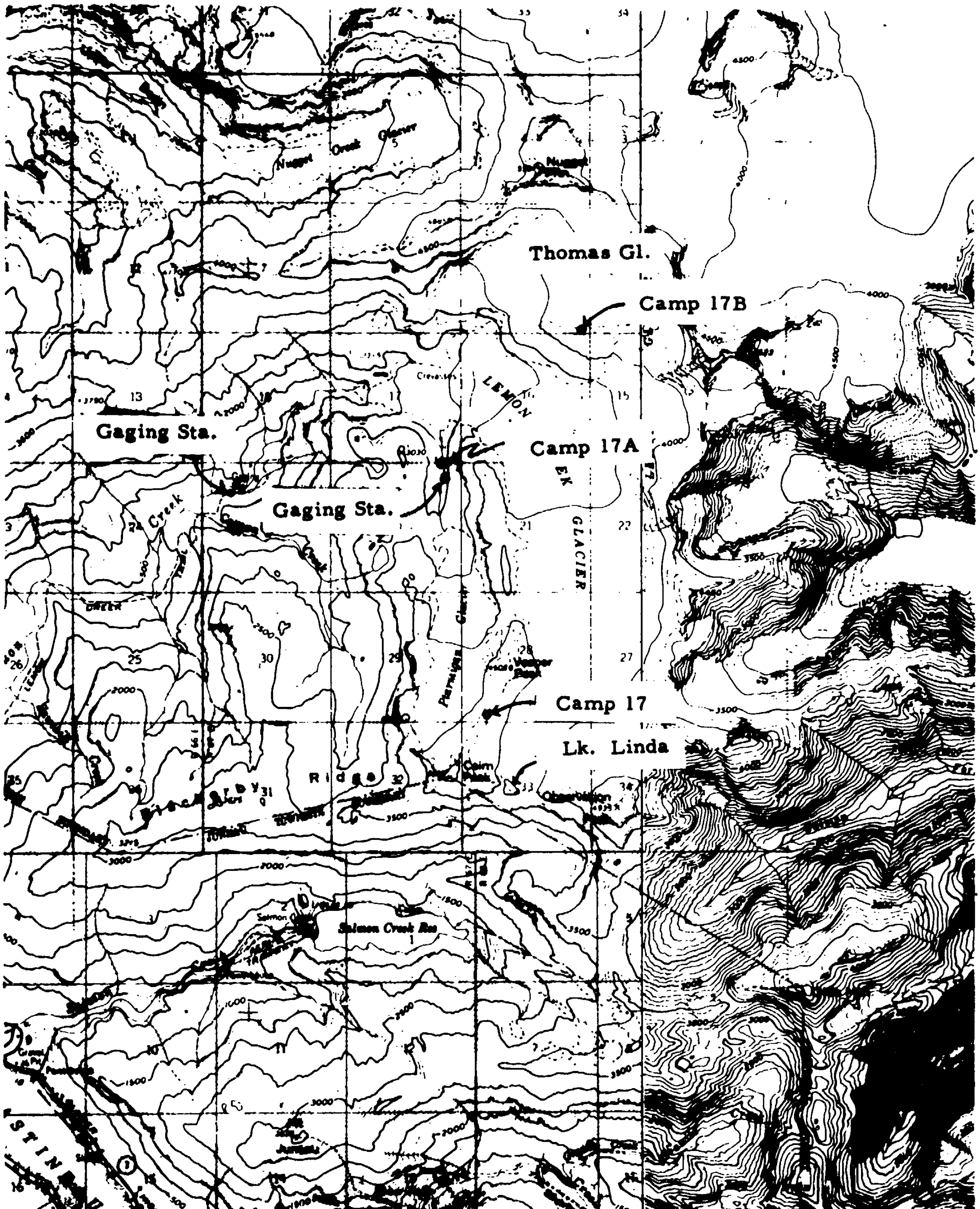
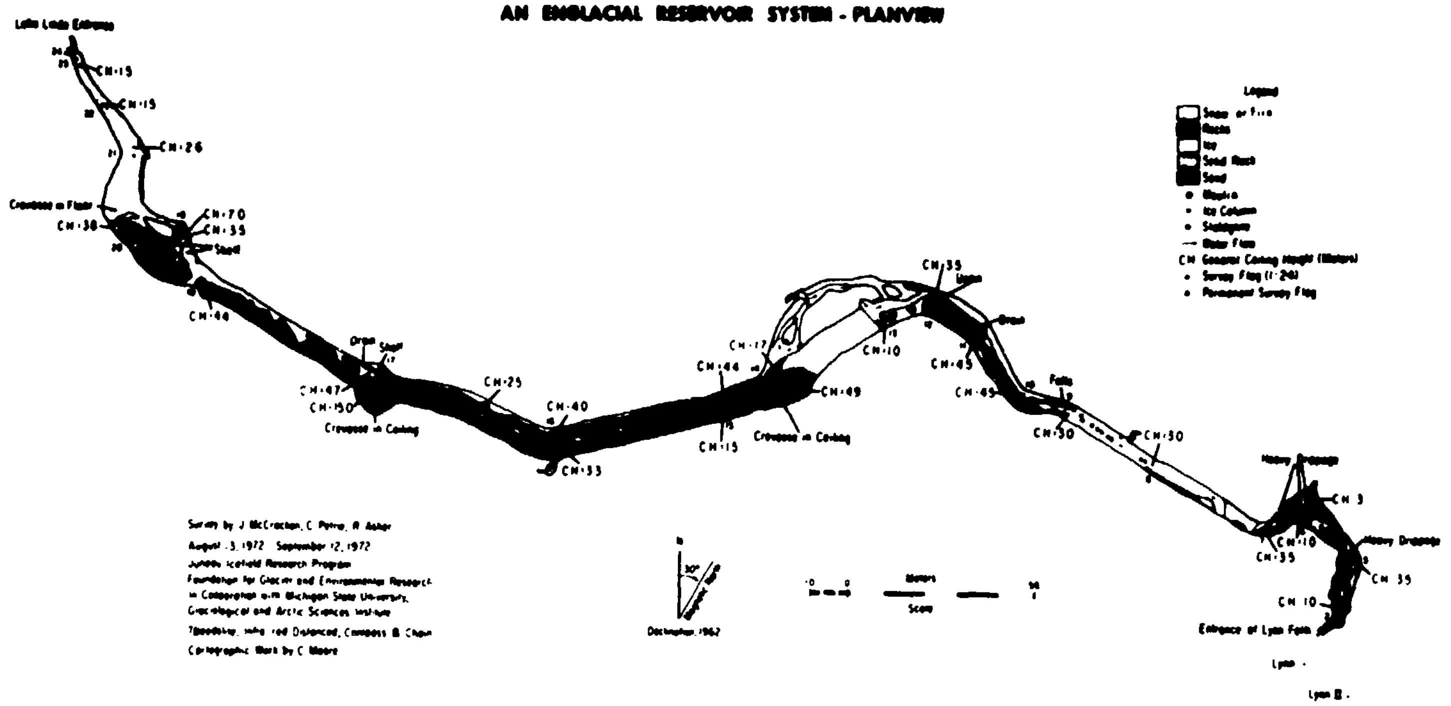


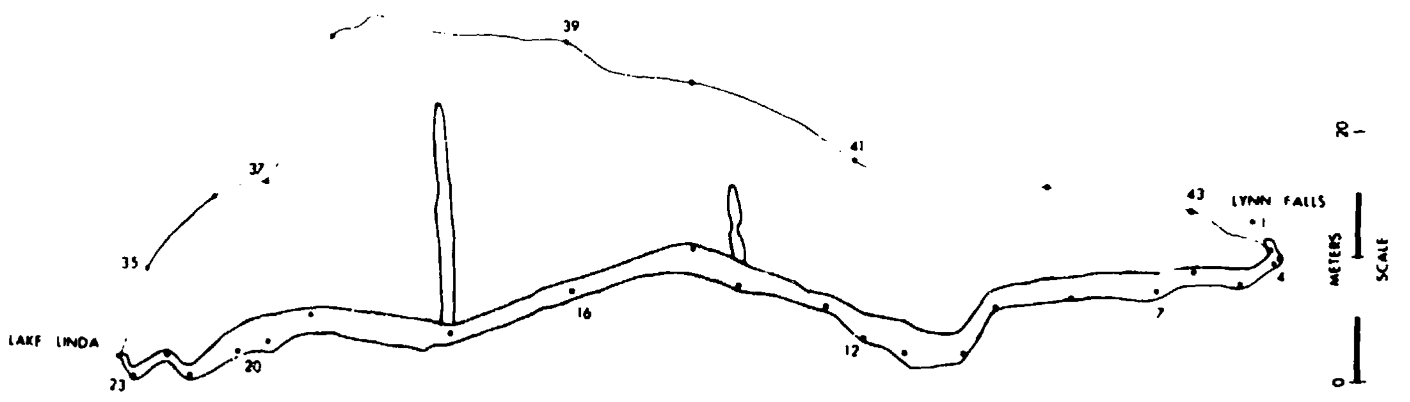
FIGURE 23 MAP OF SOUTHERN TIP OF JUNEAU ICEFIELD SHOWING LOCATION OF LEMON AND PTARMIGAN GLACIERS AND CAMPS 17, 17A AND 17B. SCALE 1:63:360

1972 GLACIER CAVE BETWEEN LAKE LINDA AND LYNN FALLS
ALONG HEADWALL MORAINE OF THE LEMON GLACIER, ALASKA, U.S.A.

AN ENGLACIAL RESERVOIR SYSTEM - PLANVIEW



AN ENGLACIAL RESERVOIR SYSTEM CROSS SECTION



SURVEY BY J McCRACKEN C PETRIE R ASHER
AUGUST 13 1972 - SEPTEMBER 12 1972
THEODOLITE INFRA RED DISTANCED COMPASS & CHAIN
JUNEAU ICEFIELD RESEARCH PROGRAM
FOUNDATION FOR GLACIER AND ENVIRONMENTAL RESEARCH
IN COOPERATION WITH MICHIGAN STATE UNIVERSITY
GLACIOLOGICAL AND ARCTIC SCIENCES INSTITUTE
CARTOGRAPHIC WORK BY C MOORE

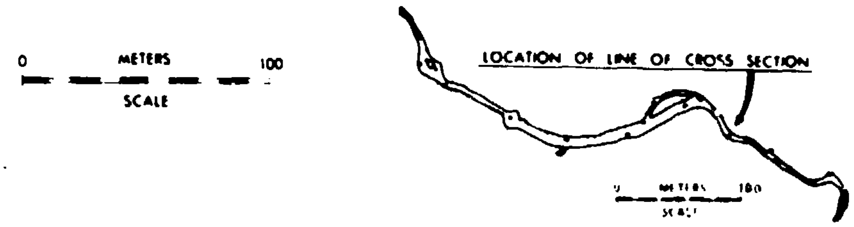


Fig. 24

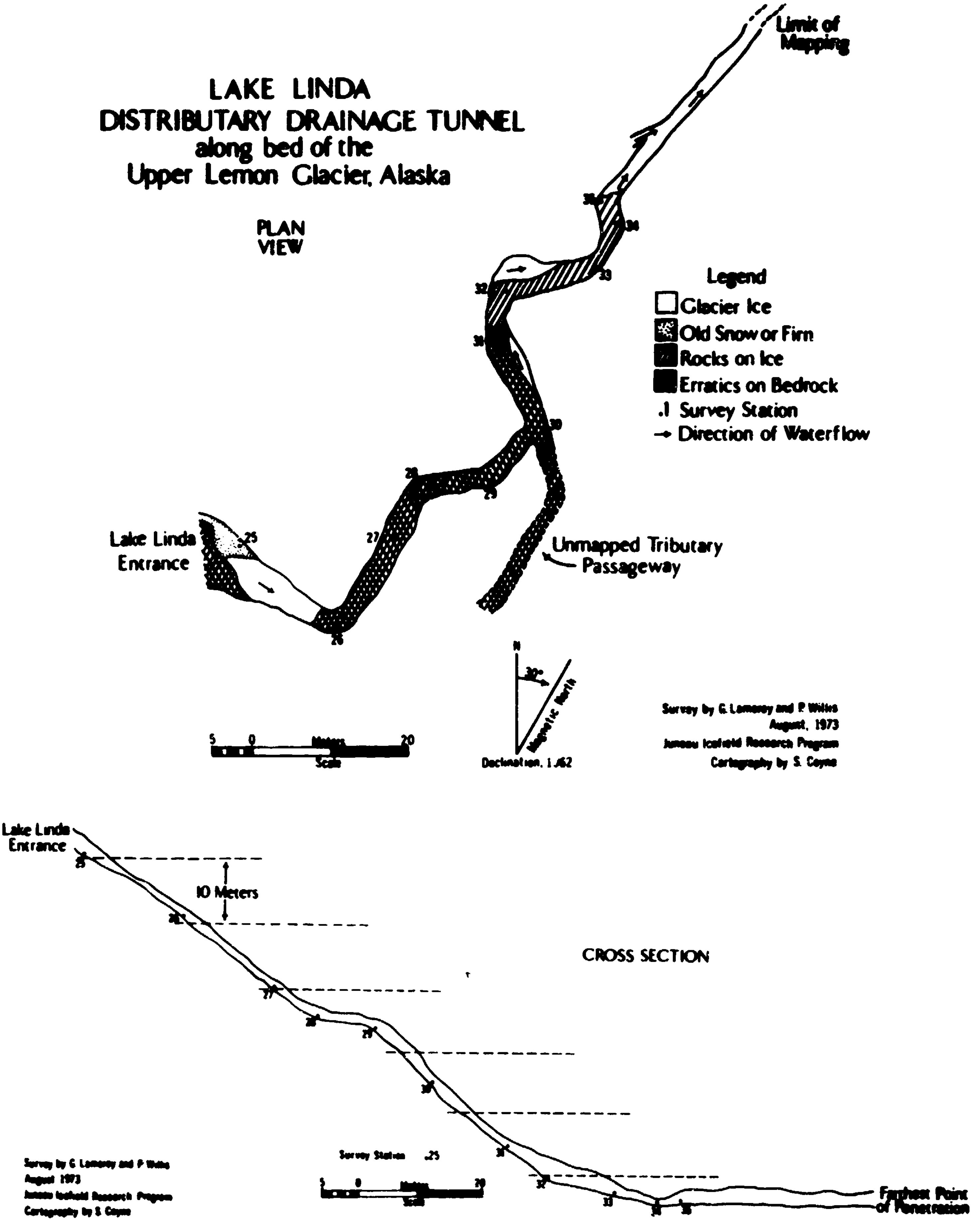
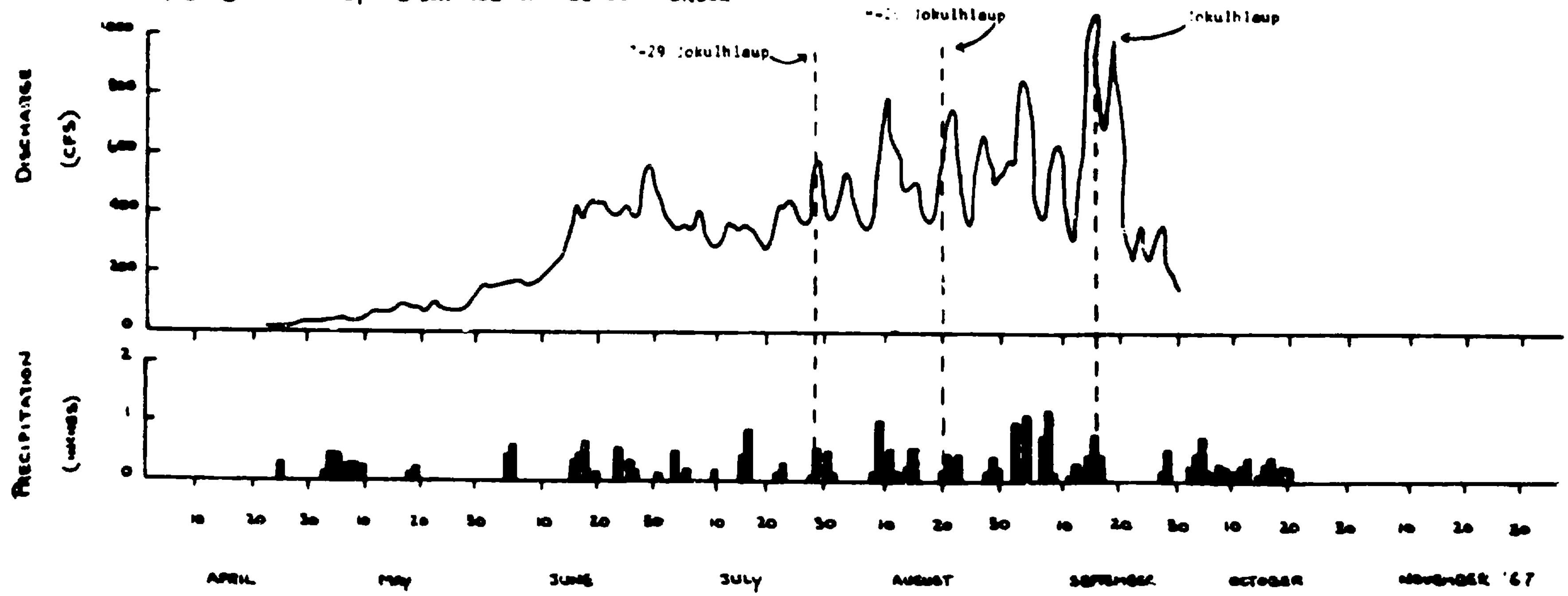

LAKE LINDA
DISTRIBUTARY DRAINAGE TUNNEL
along bed of the
Upper Lemon Glacier, Alaska
PLAN
VIEW
Limit of
Mapping
Legend
Glacier Ice
Old Snow or Firn
Rocks on Ice
Erratics on Bedrock
.1 Survey Station
→ Direction of Waterflow
Lake Linda
Entrance
Unmapped Tributary
Passageway
N
30°
Magnetic North
Survey by G. Lamorey and P. Willis
August, 1973
Juneau Icefield Research Program
Cartography by S. Coyne
5 0 Meters 20
Scale
Lake Linda
Entrance
10 Meters
CROSS SECTION
Survey Station .25
Farthest Point
of Penetration

Fig. 25

Fig 26 Daily Discharge of Le[illegible] Creek



Daily Precipitation (≥ 0.1") at Juneau Air[illegible]t

FIG. 27a RELATIVE HUMIDITY AND DEW POINT VS. TIME AT CAMP 17

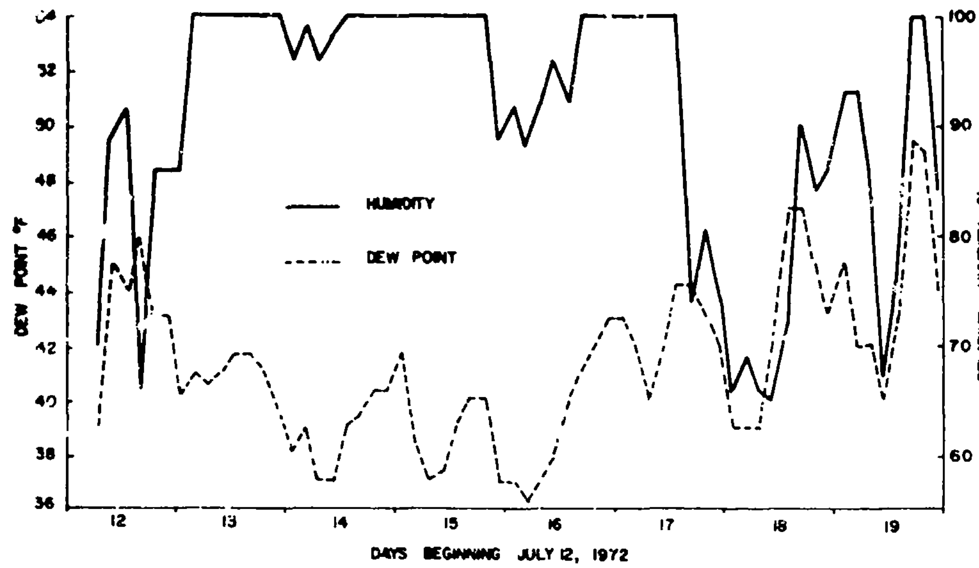


FIG. 27b CAMP 17 TEMPERATURES, RADIATION, AND CLOUD COVER OVER A FEW DAYS IN JULY 1972

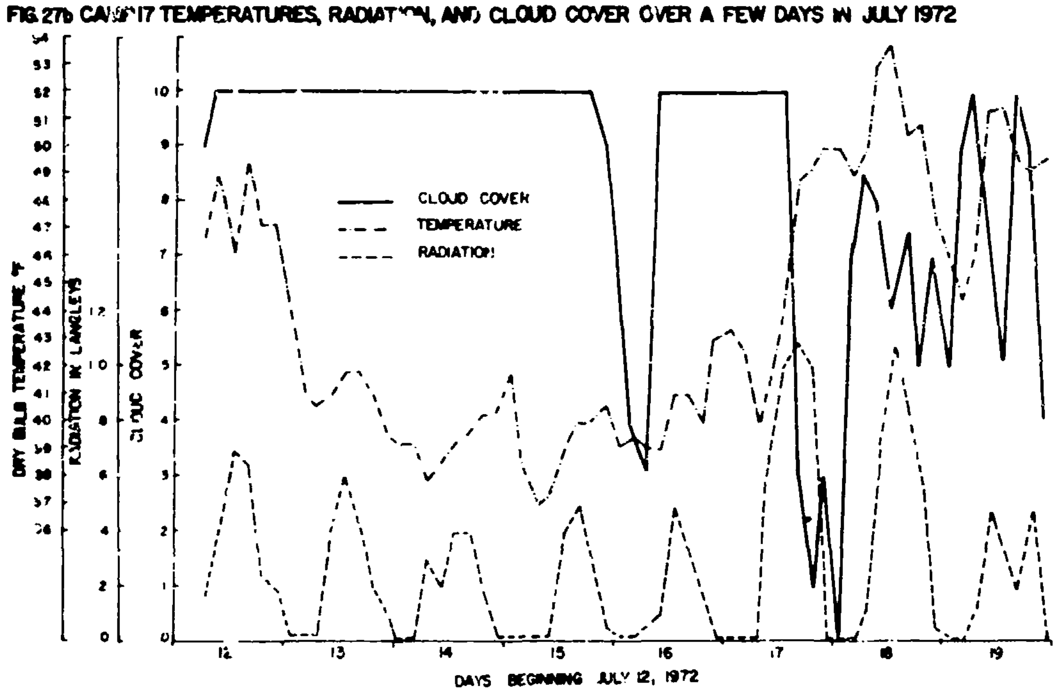


FIG. 27c ABLATION AND WATER EQUIVALENT ON FIRN AT CAMP 17 (4,200 ft.)

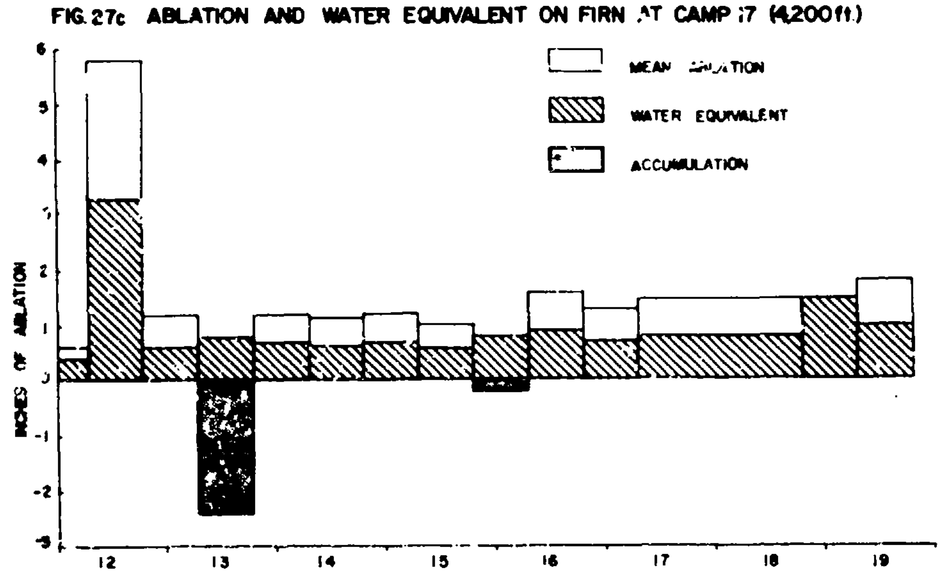
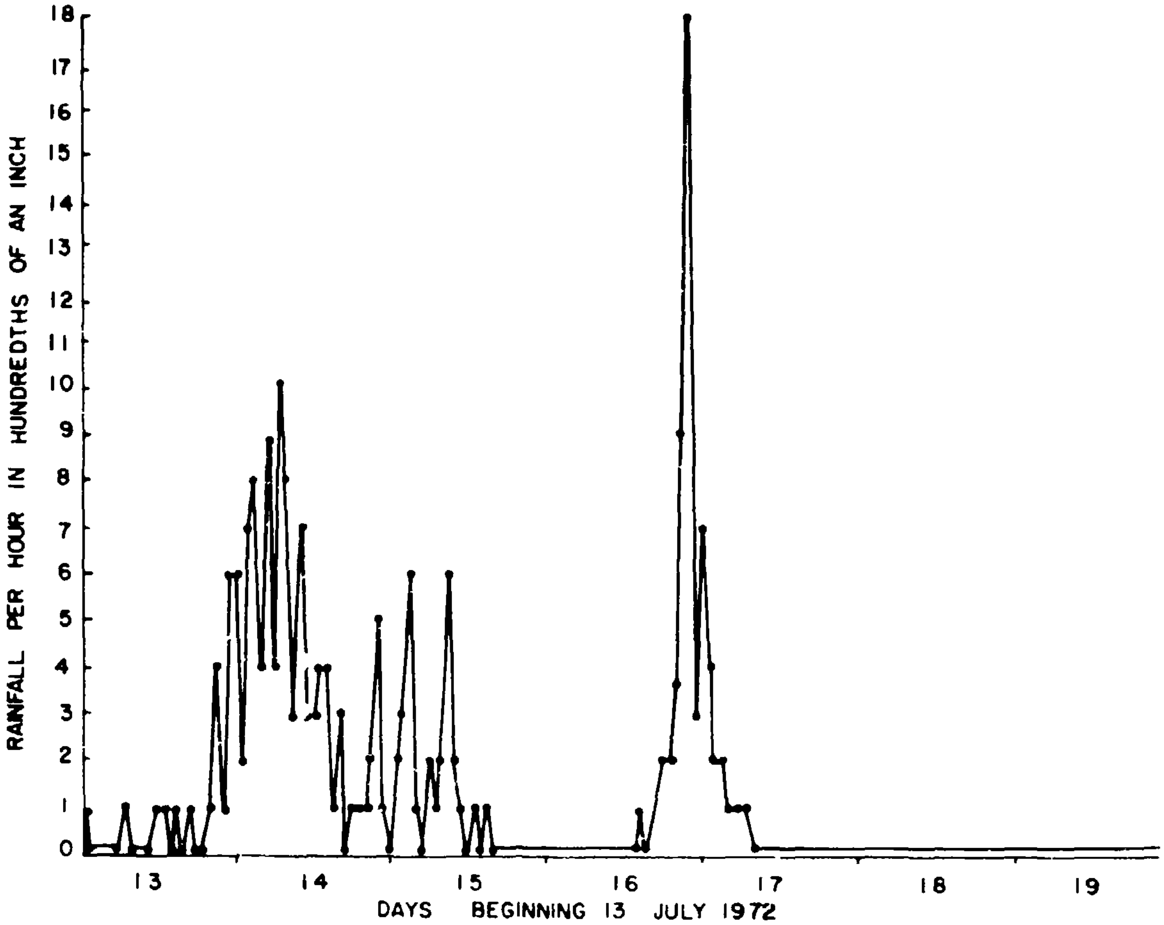


FIG. 28a - PRECIPITATION FROM MRI AT CAMP 17, LEMON GLACIER



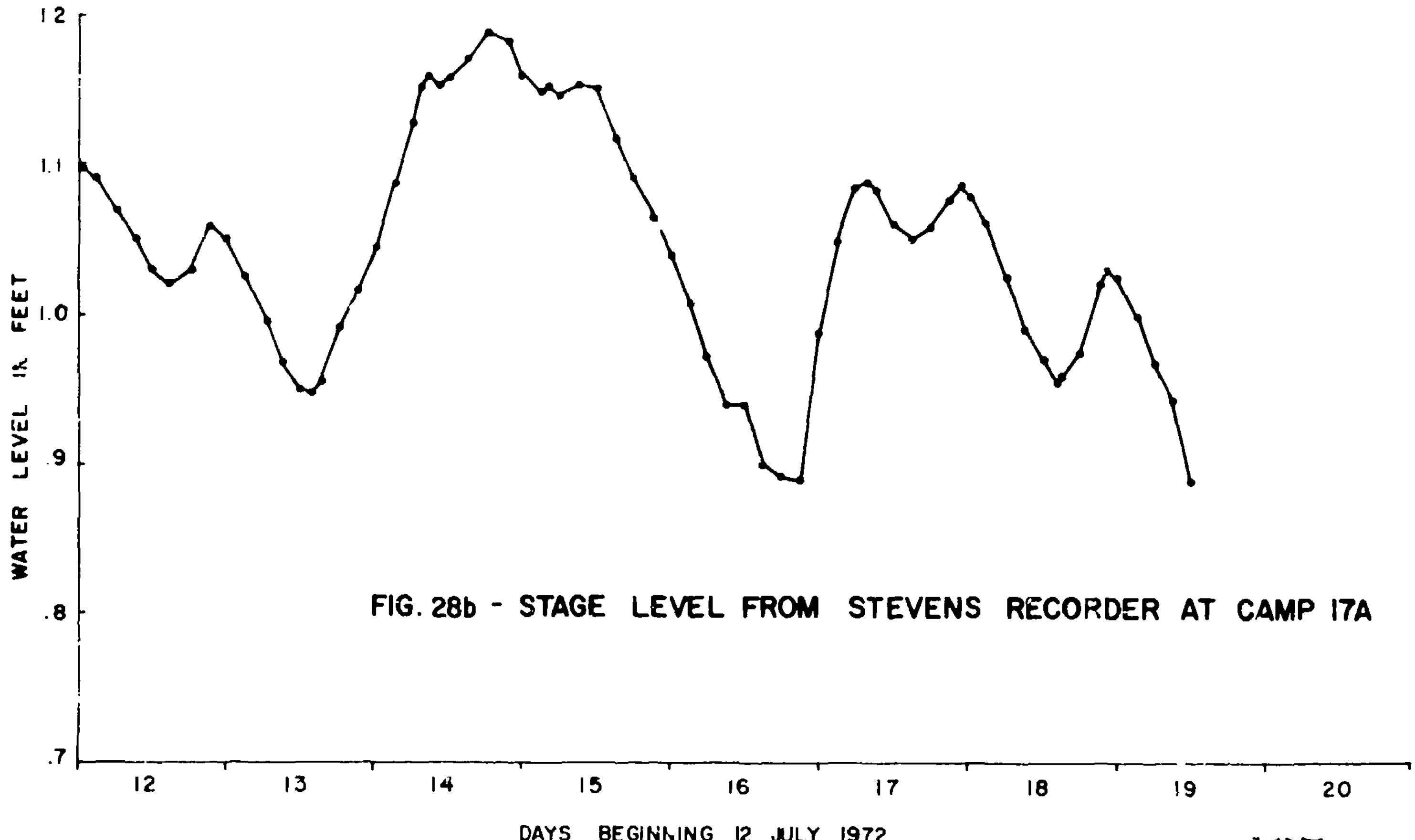
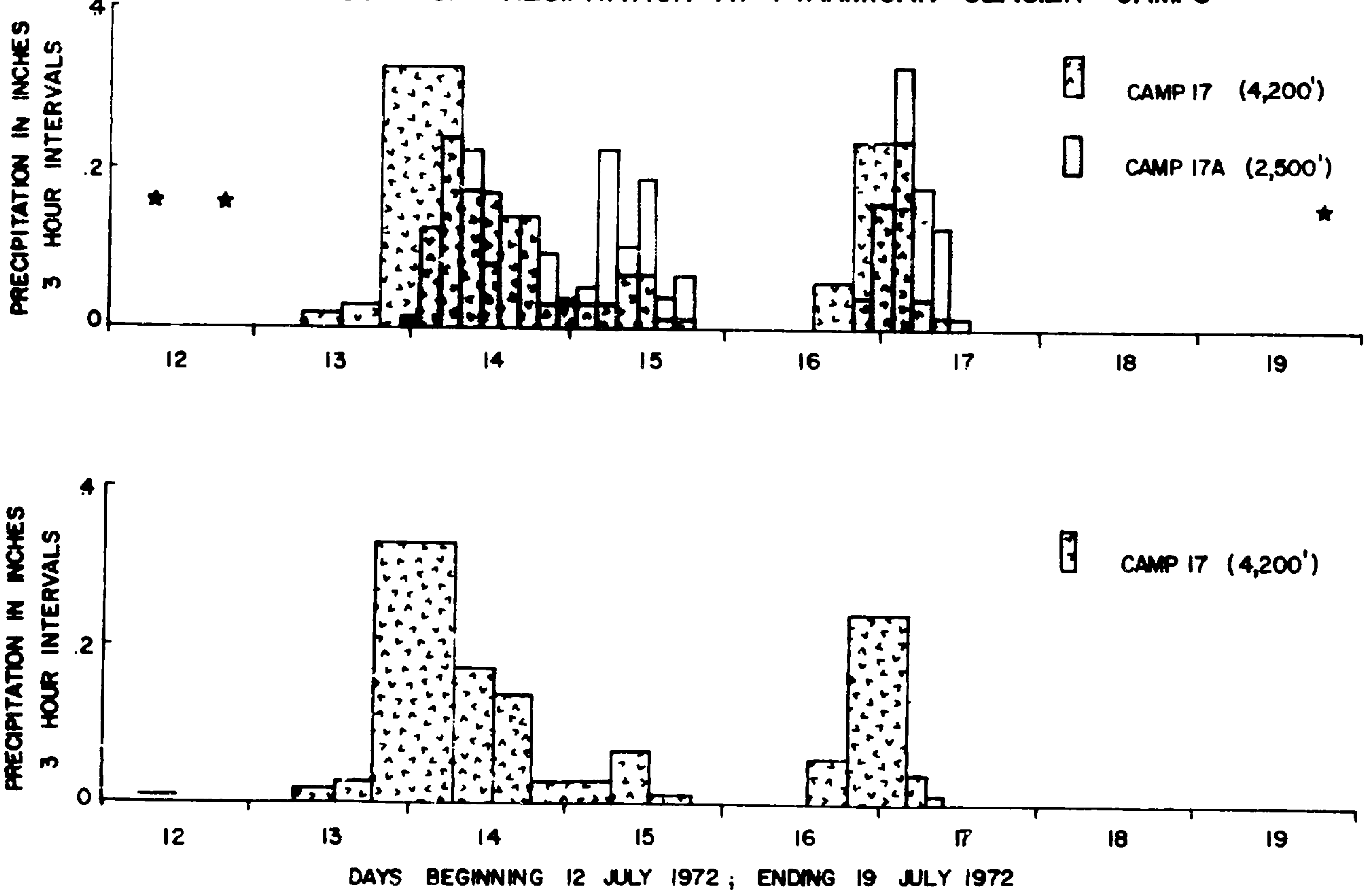


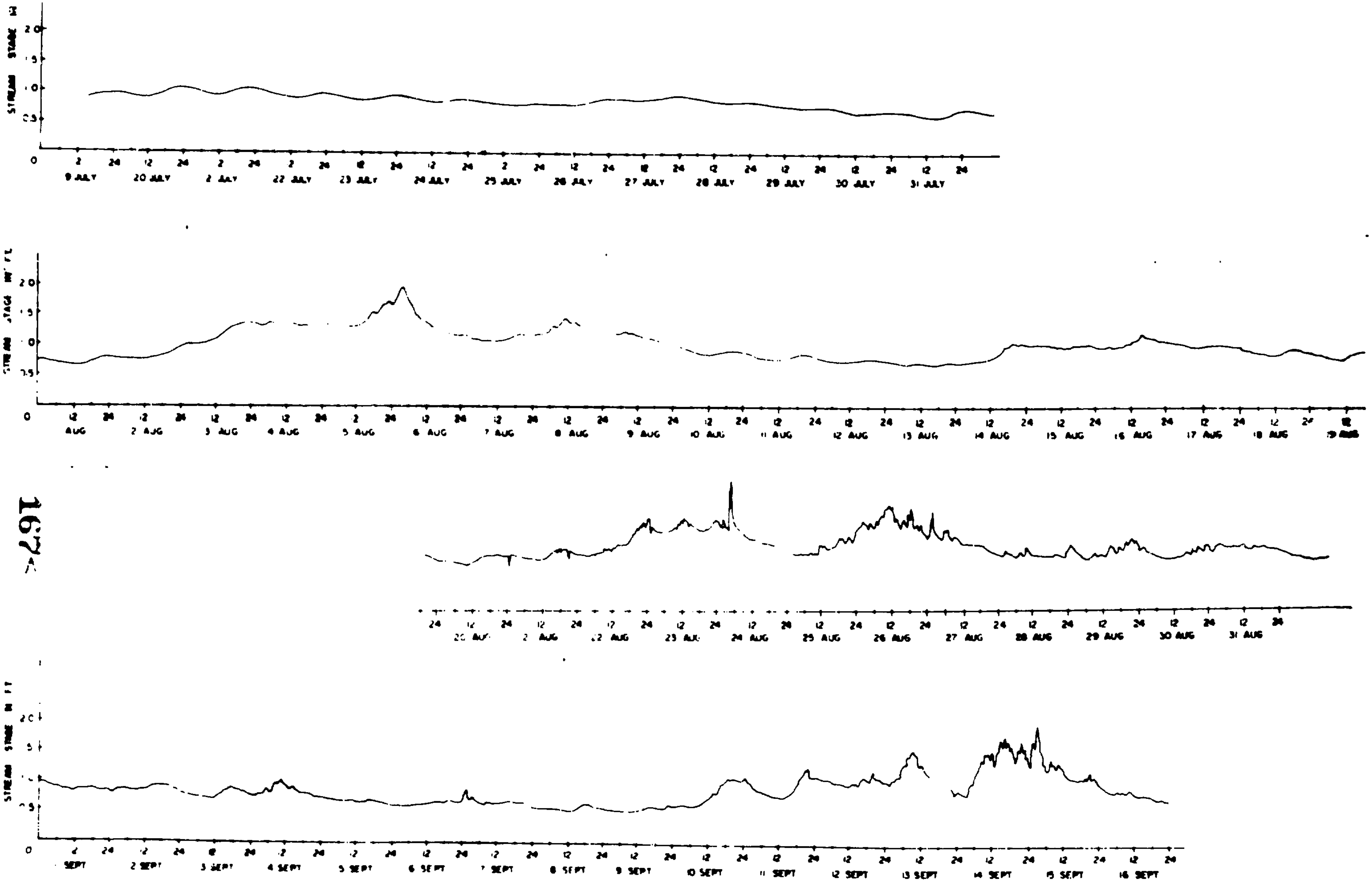
FIG. 28b - STAGE LEVEL FROM STEVENS RECORDER AT CAMP 17A

FIG. 29 COMPARISON OF PRECIPITATION AT PTARMIGAN GLACIER CAMPS



★ Data Not Available

FIG-30 STREAM STAGE OF THE PTARMIGAN CREEK VERSUS TIME - SUMMER 1972



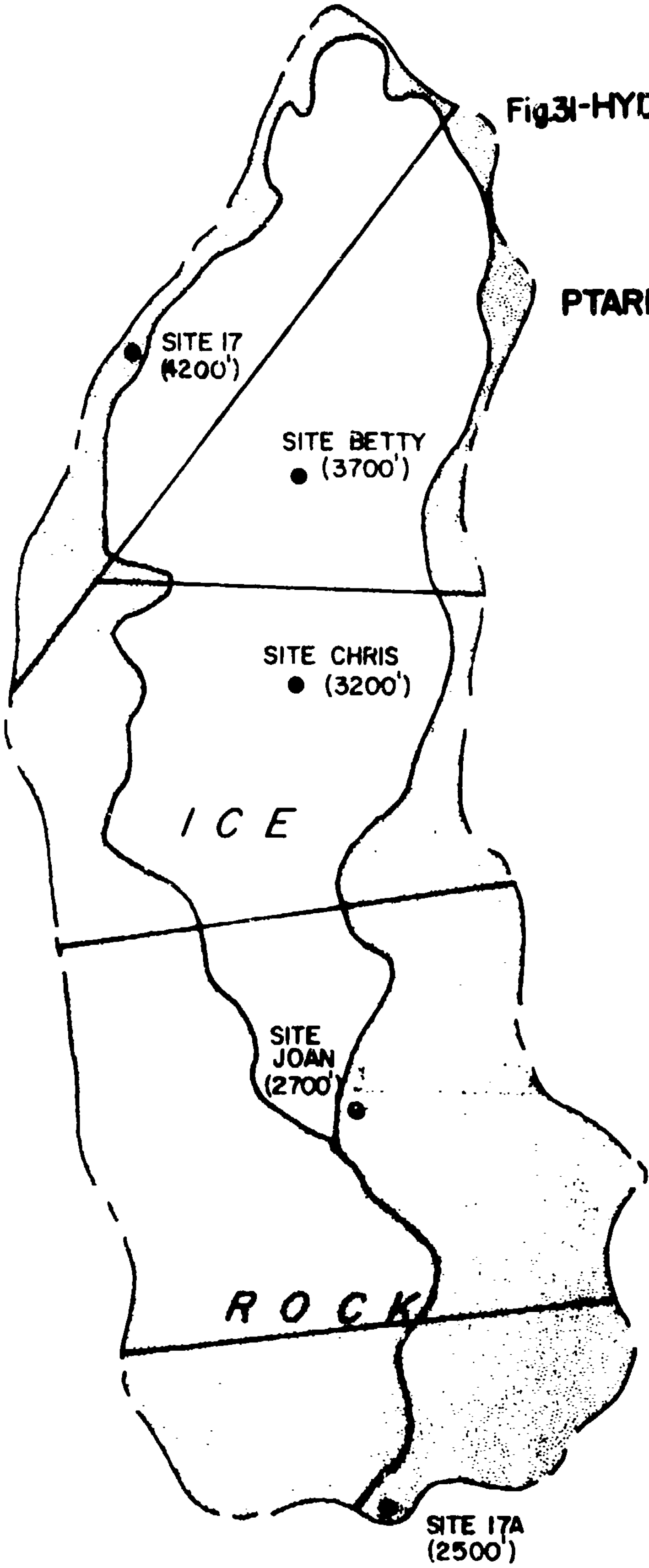


Fig.31-HYDROLOGIC SECTORS OF THE PTARMIGAN GLACIER BASIN

AREAS IN $(METER)^2$
TOTAL BASIN 3.9×10^6
TOTAL GLACIER 1.7×10^6

C-17 SECTOR
GLACIER 3.5×10^5
BASIN 5.2×10^5

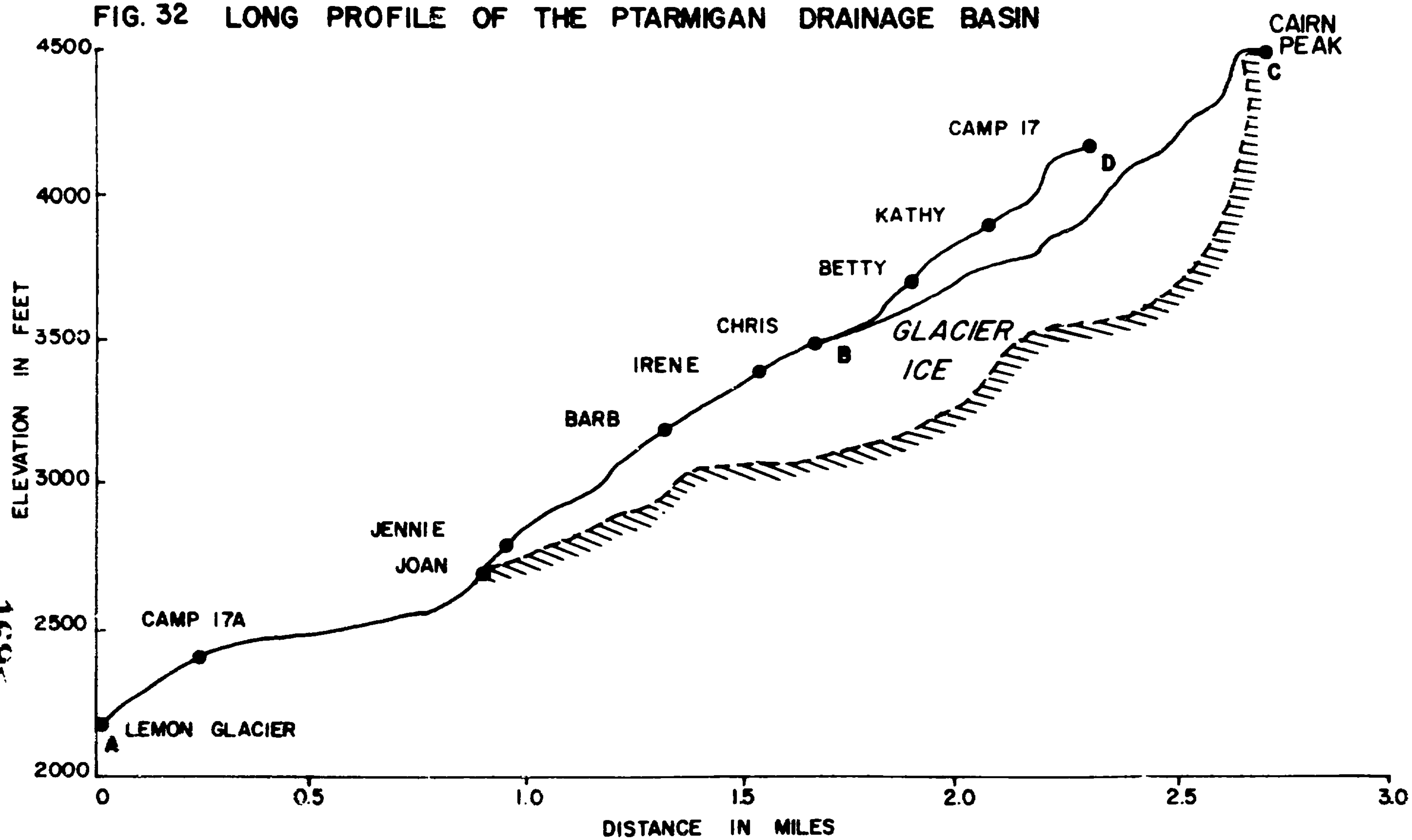
BETTY SECTOR
GLACIER 5.7×10^5
BASIN 6.7×10^5

CHRIS SECTOR
GLACIER 5.7×10^5
BASIN 9.5×10^5

JOAN SECTOR
GLACIER 1.7×10^5
BASIN 12.8×10^5

C-17A SECTOR
BASIN 5.2×10^5

FIG. 32 LONG PROFILE OF THE PTARMIGAN DRAINAGE BASIN



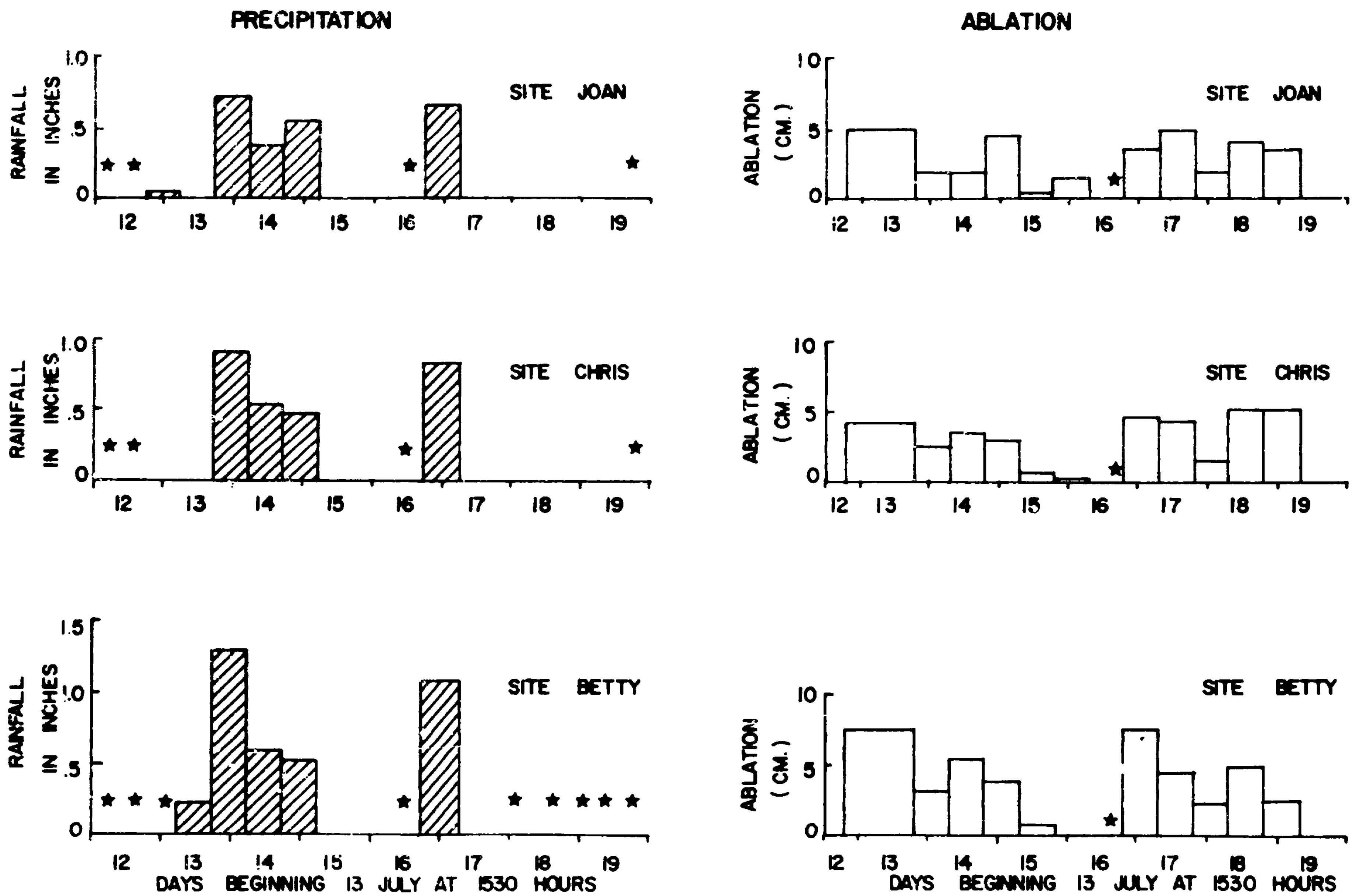


FIG. 33 COMPARATIVE ABLATION ON PTARMIGAN GLACIER, JULY 1972

★ Data Not Available

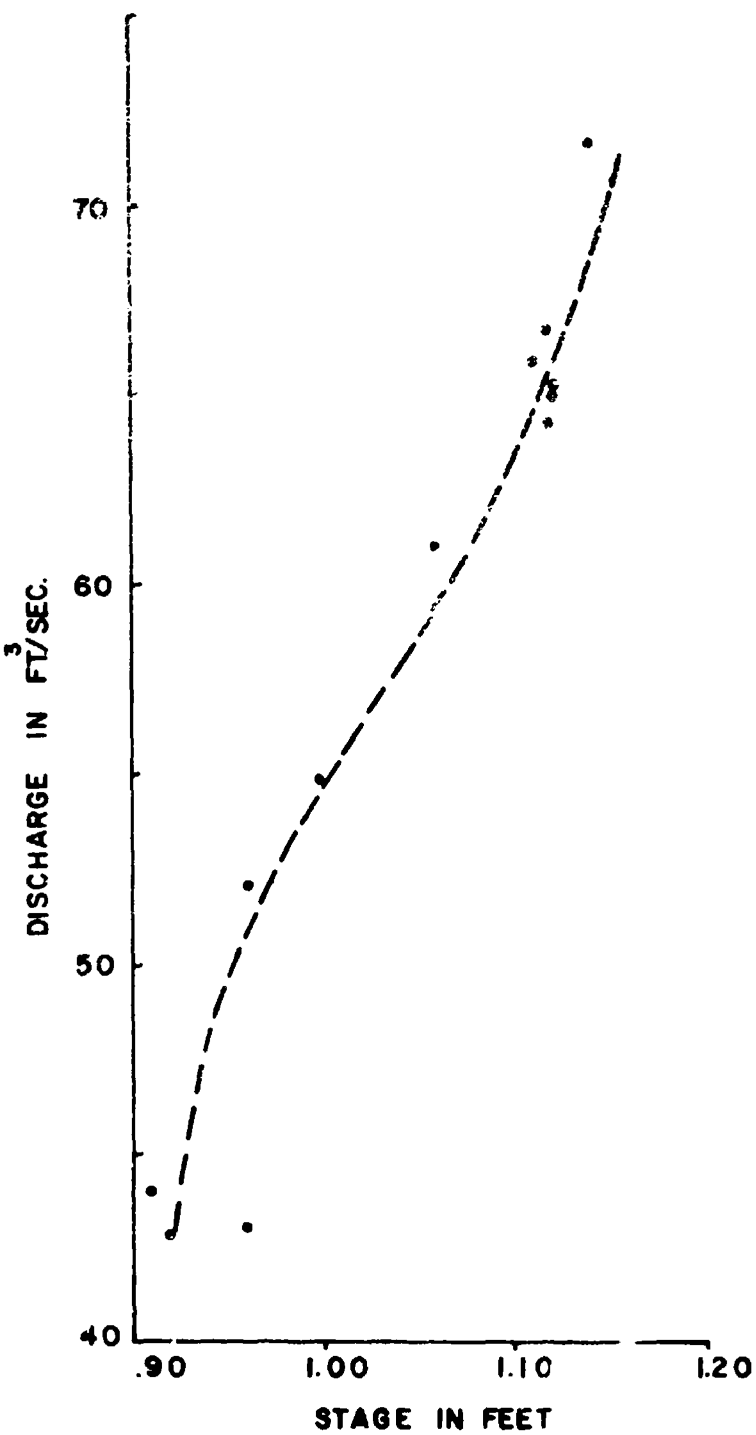
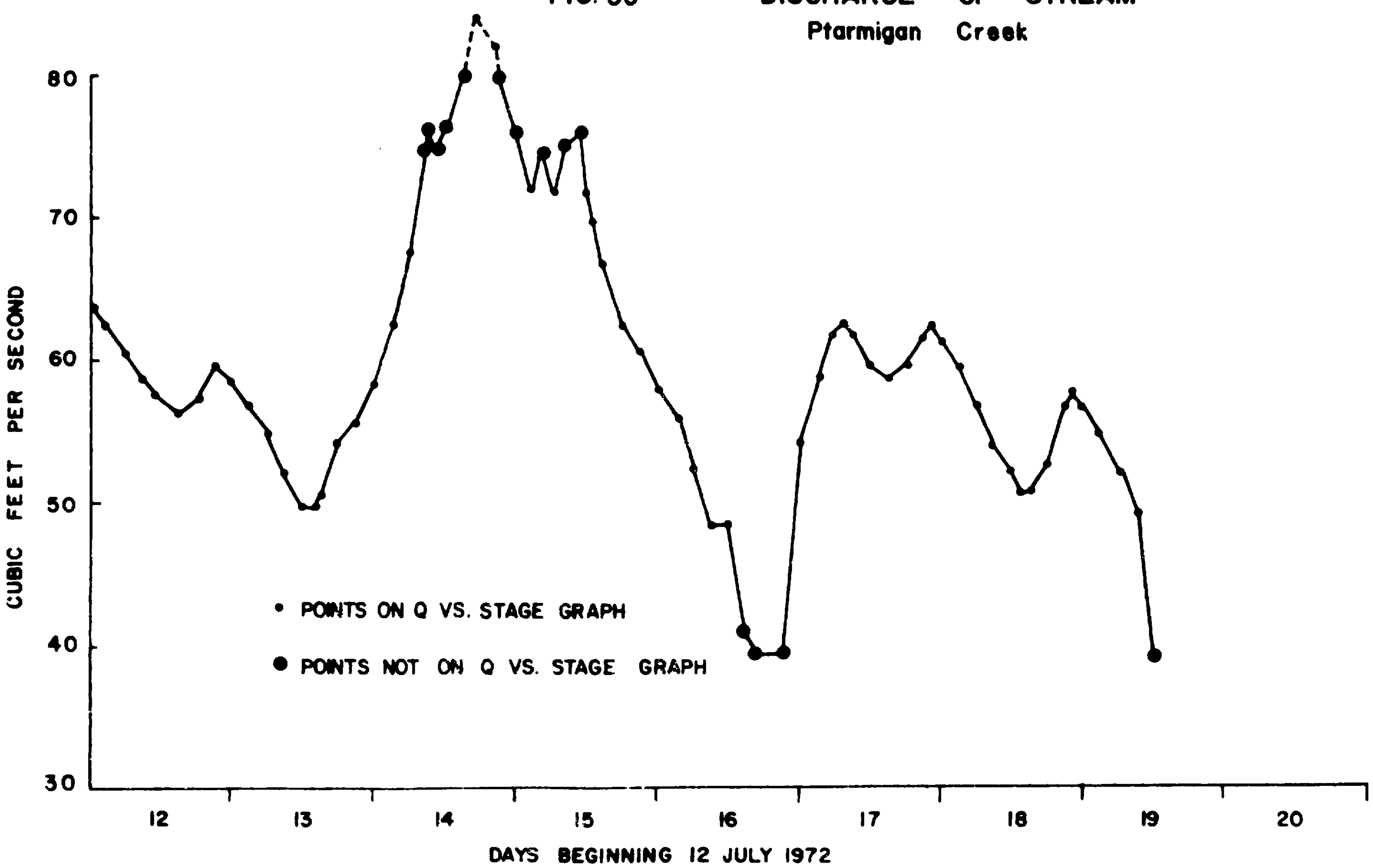


FIG. 34 — DISCHARGE (Q) VS STAGE
Ptarmigan Creek

FIG. 35 DISCHARGE OF STREAM
Ptarmigan Creek



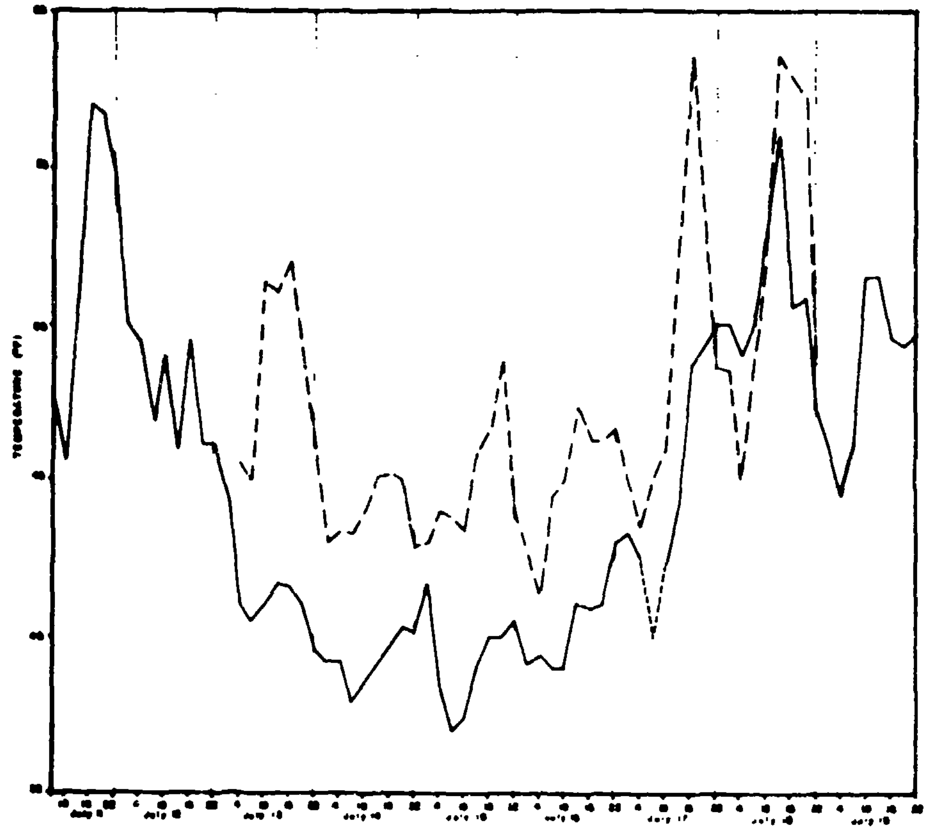
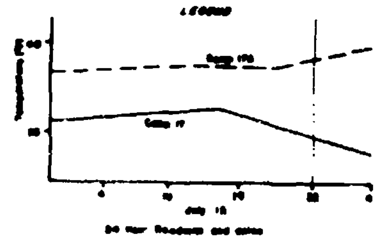


Fig.36 AMBIENT TEMPERATURE COMPARISONS
AT CAMPS 17 AND 17A
SUMMER, 1972



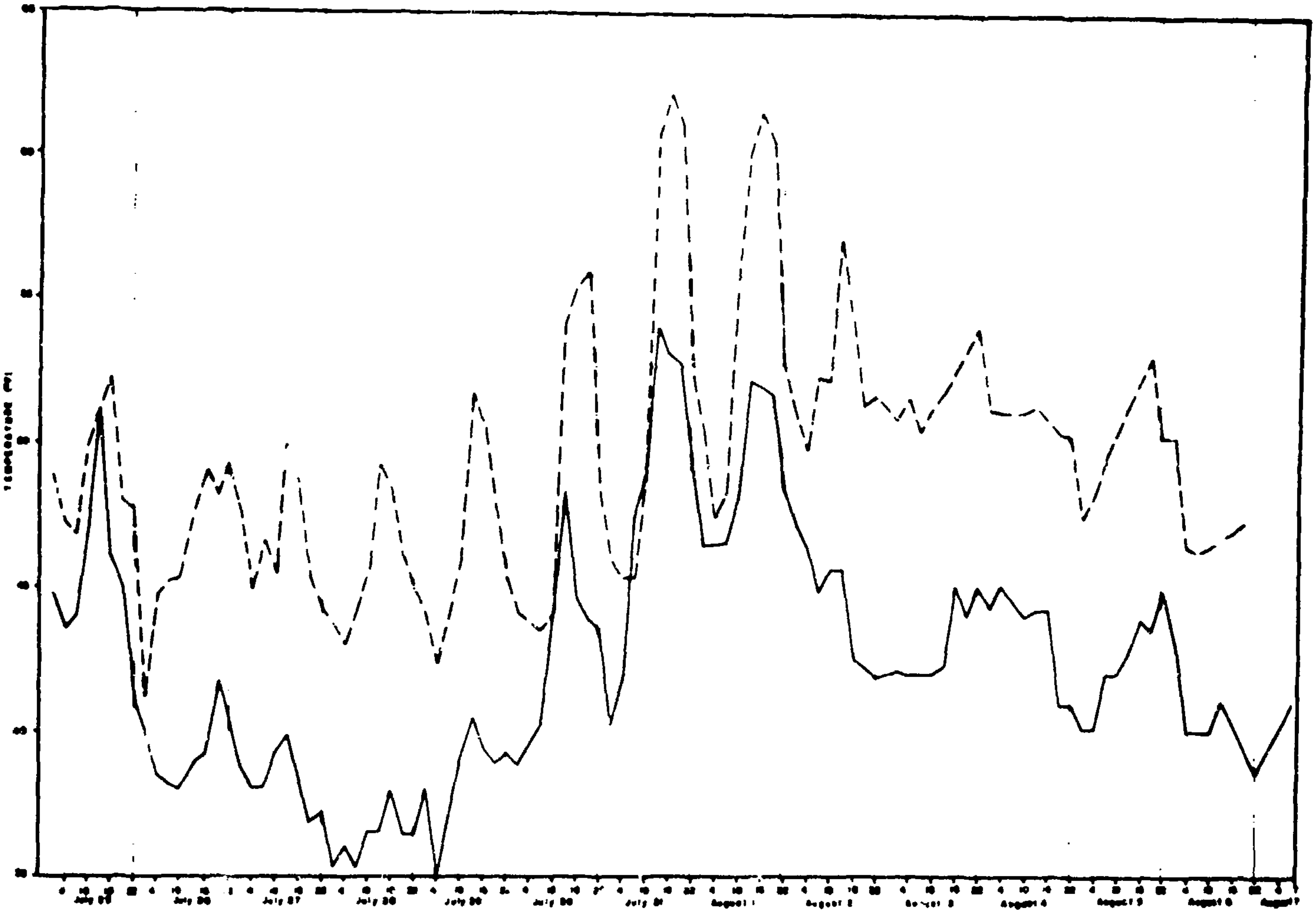
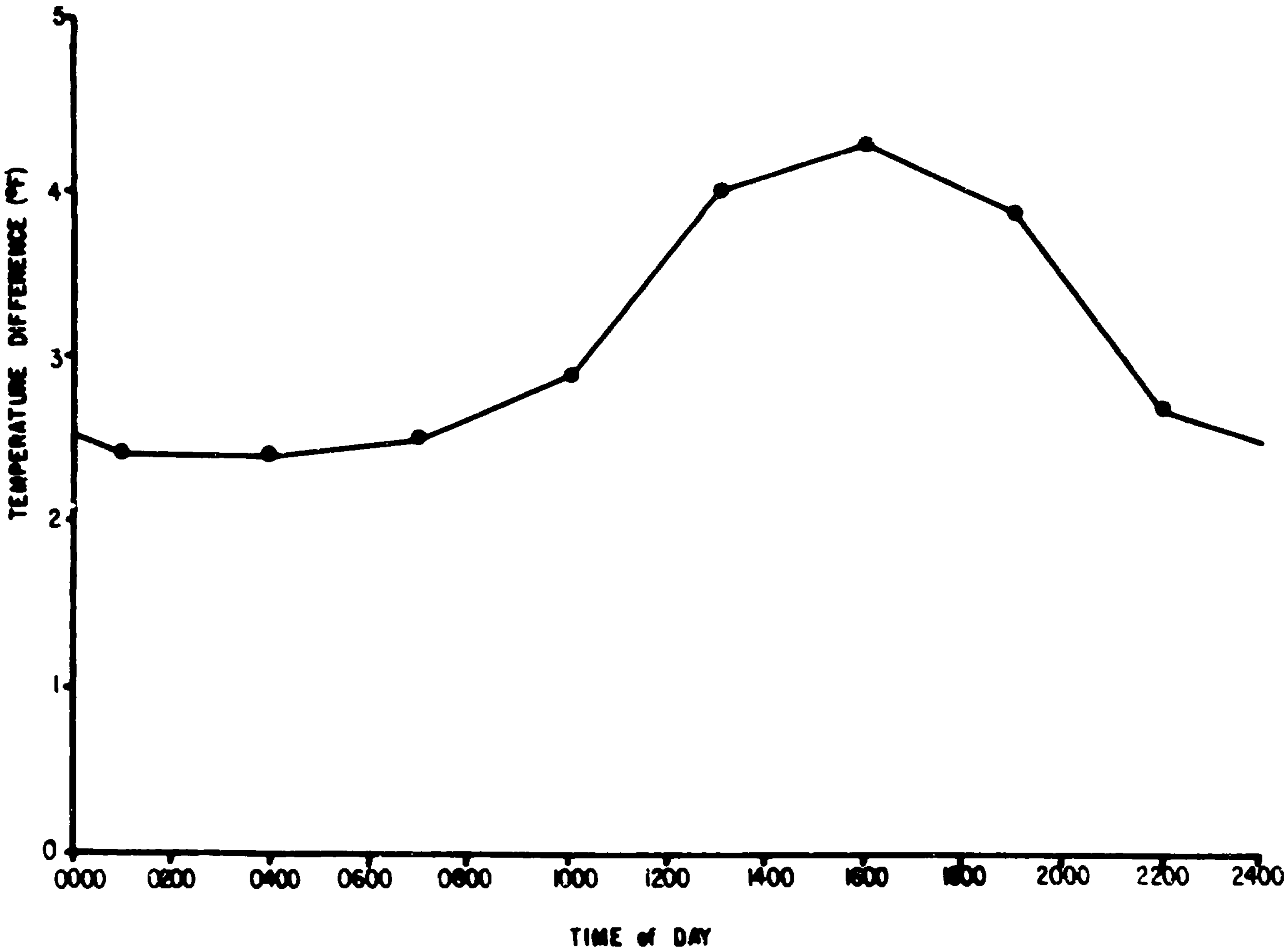


Fig. 37 - Temperature difference between Camps 17 and 17A versus time of day.



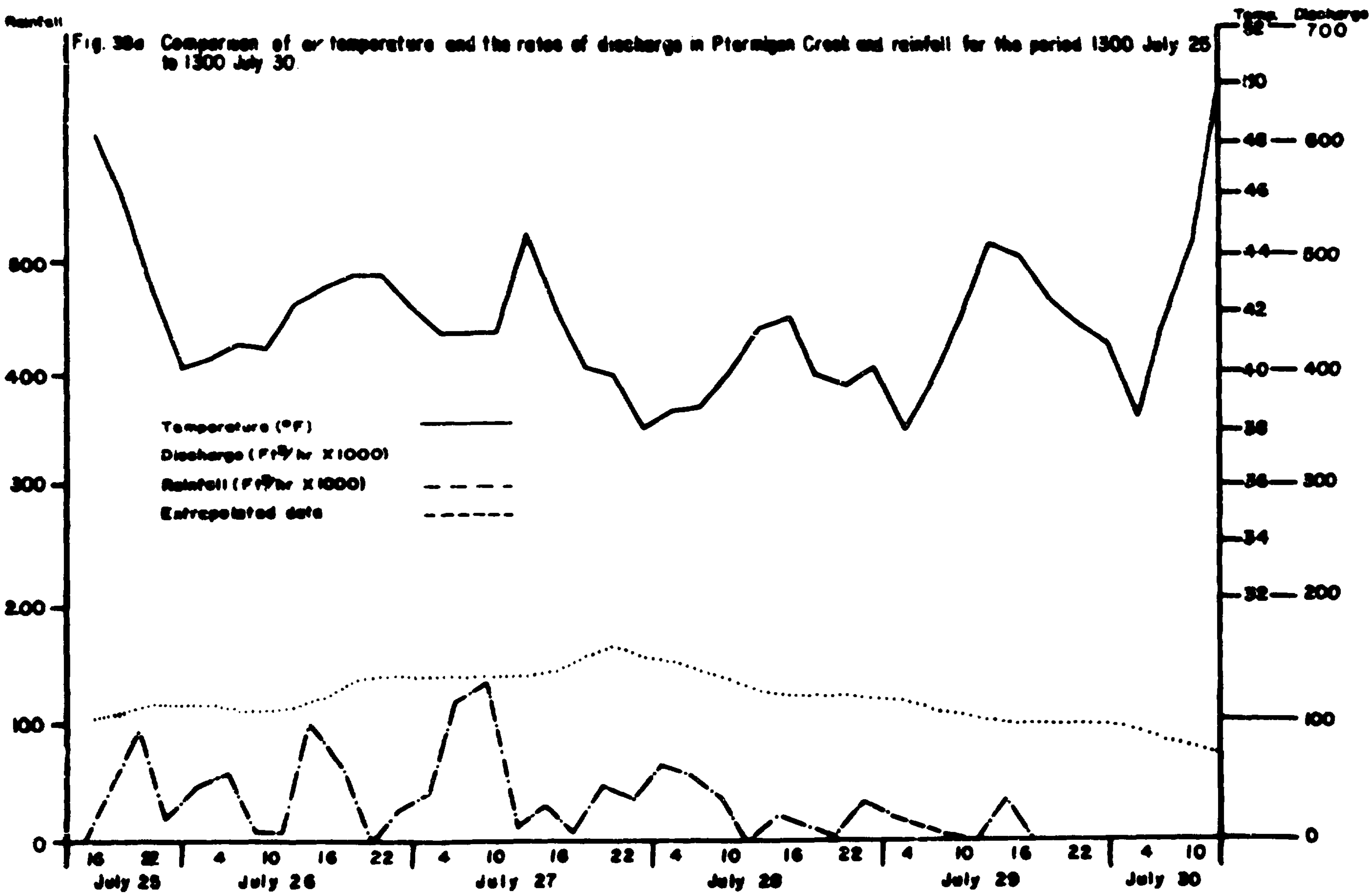
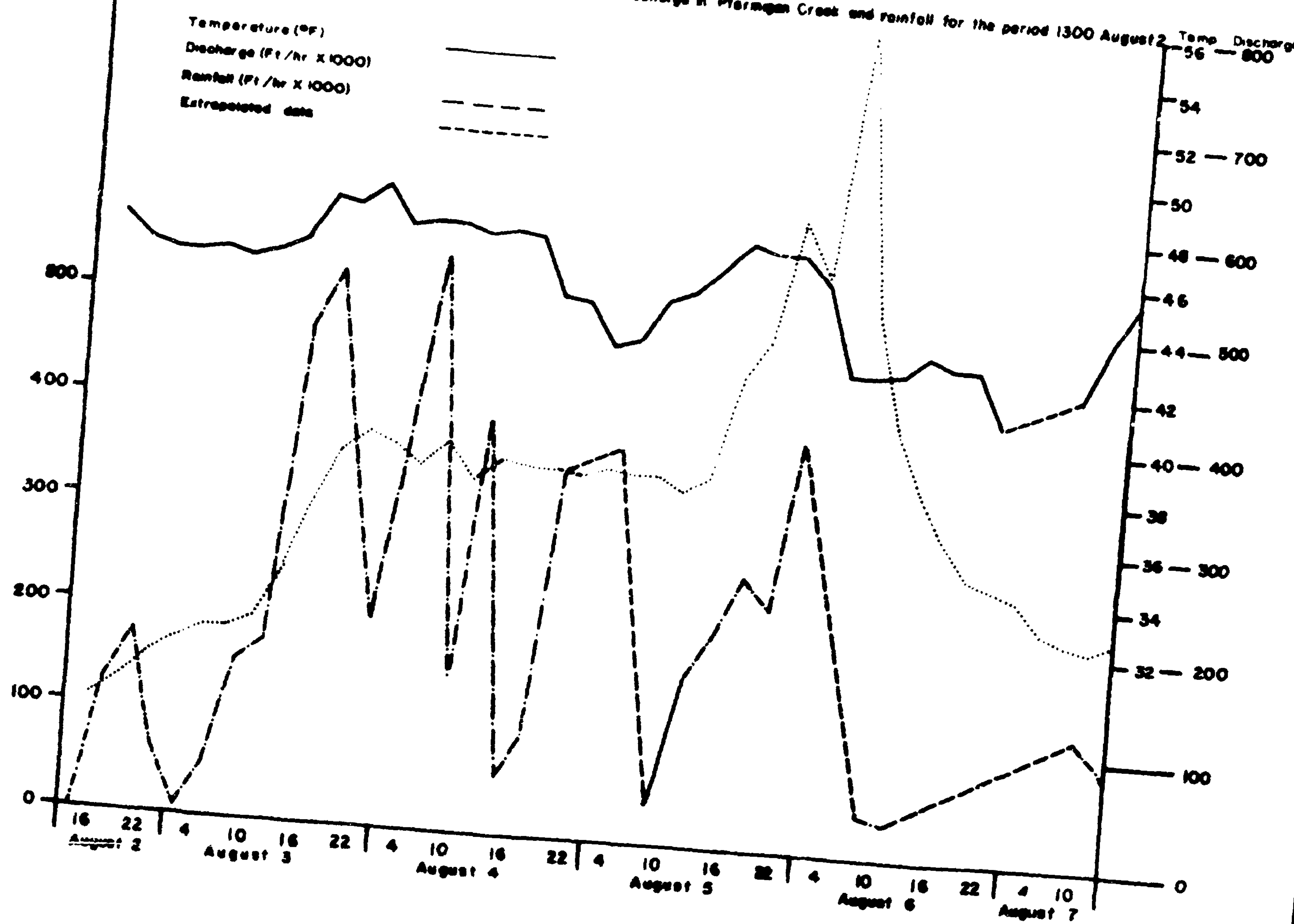


Fig. 38a Comparison of air temperature and the rates of discharge in Ptarmigan Creek and rainfall for the period 1300 July 25 to 1300 July 30

Fig 38b Comparisons of air temperature and the rates of discharge in Ptarmigan Creek and rainfall for the period 1300 August 2 to 1300 August 7



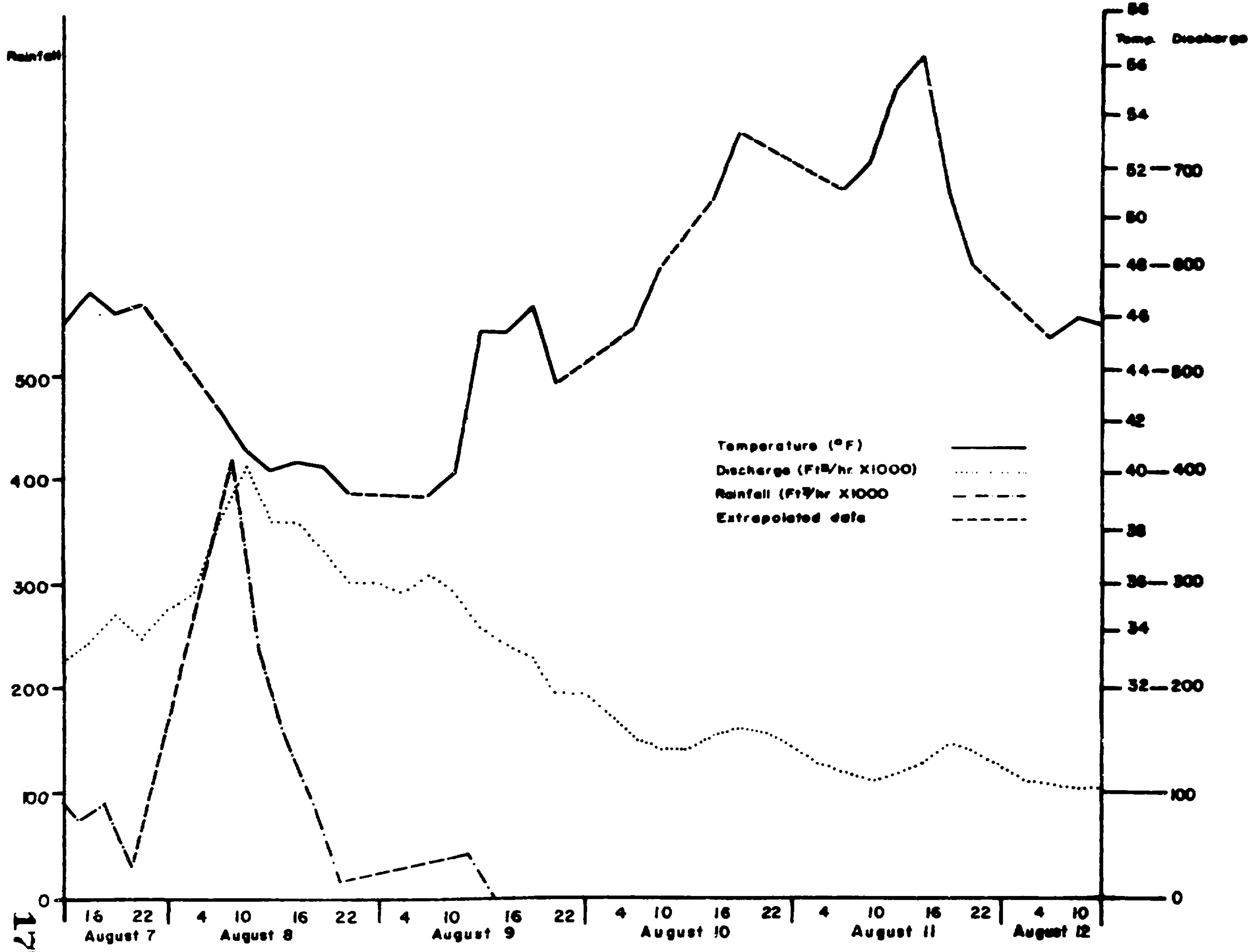
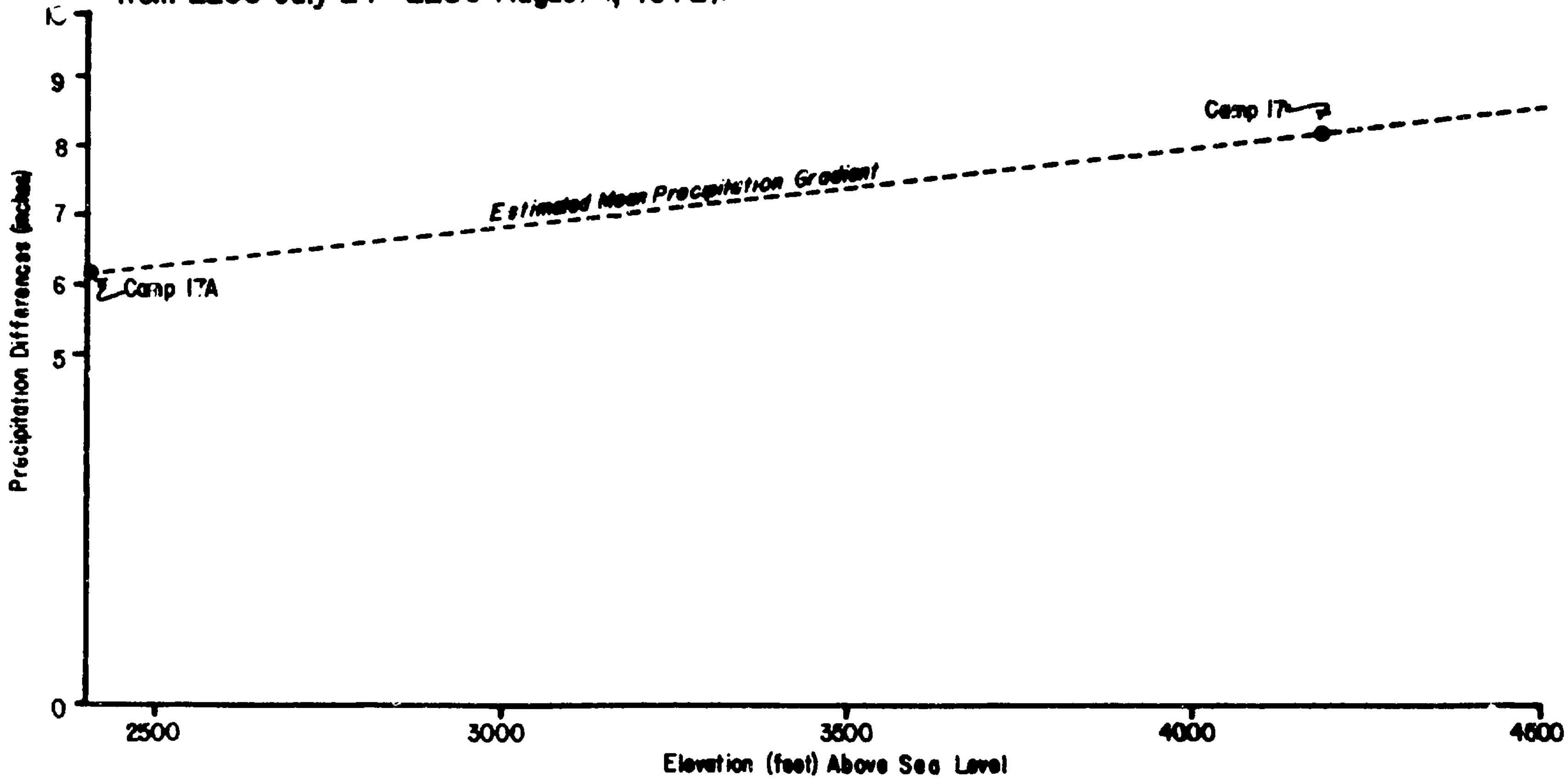
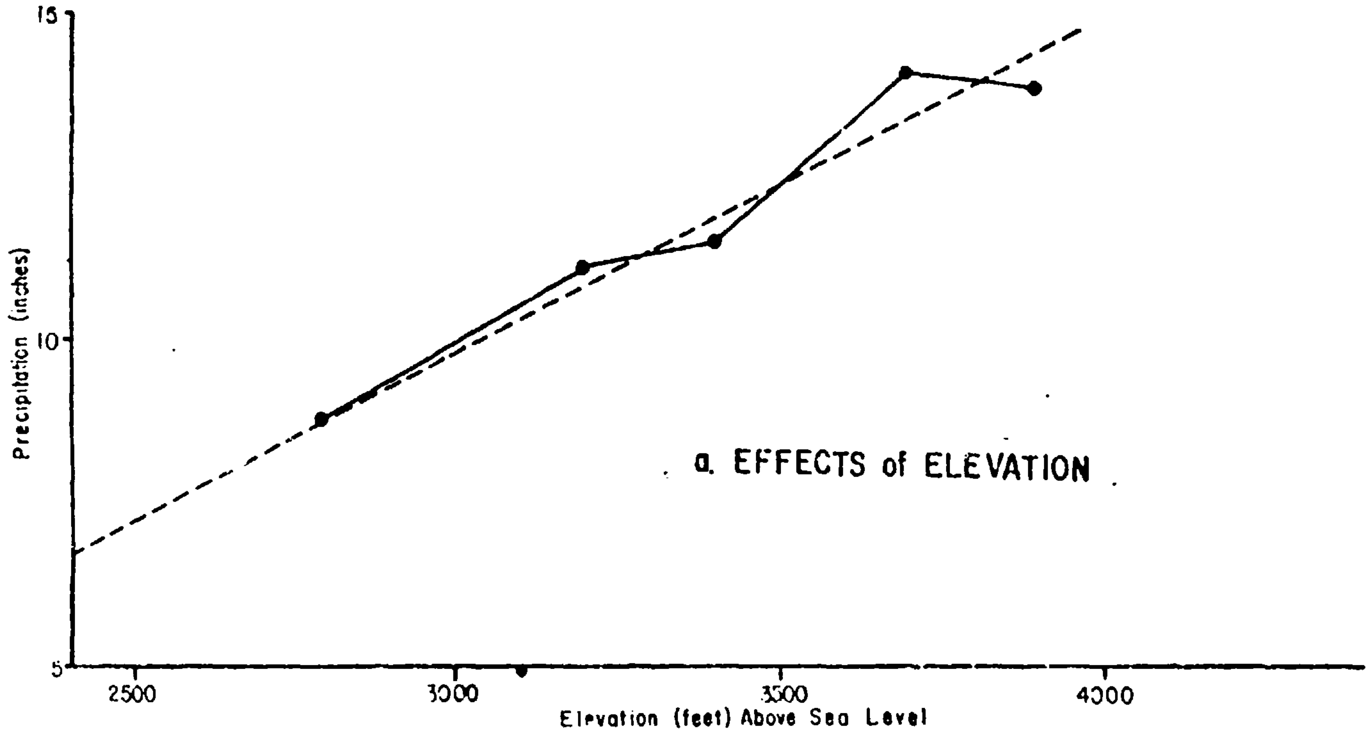


Fig.38c Comparison of air temperature and the rate of discharge in Ptamigan Creek and rainfall for the period 1300 August 7 to 1300 August 12.

Fig. 39 - Precipitation differences between Camps 17 and 17A (for the periods 2200 July 12 - 2200 July 18 and from 2200 July 24 - 2200 August 4, 1972).





Field Site....................●

Mean Precipitation Gradient....---- --

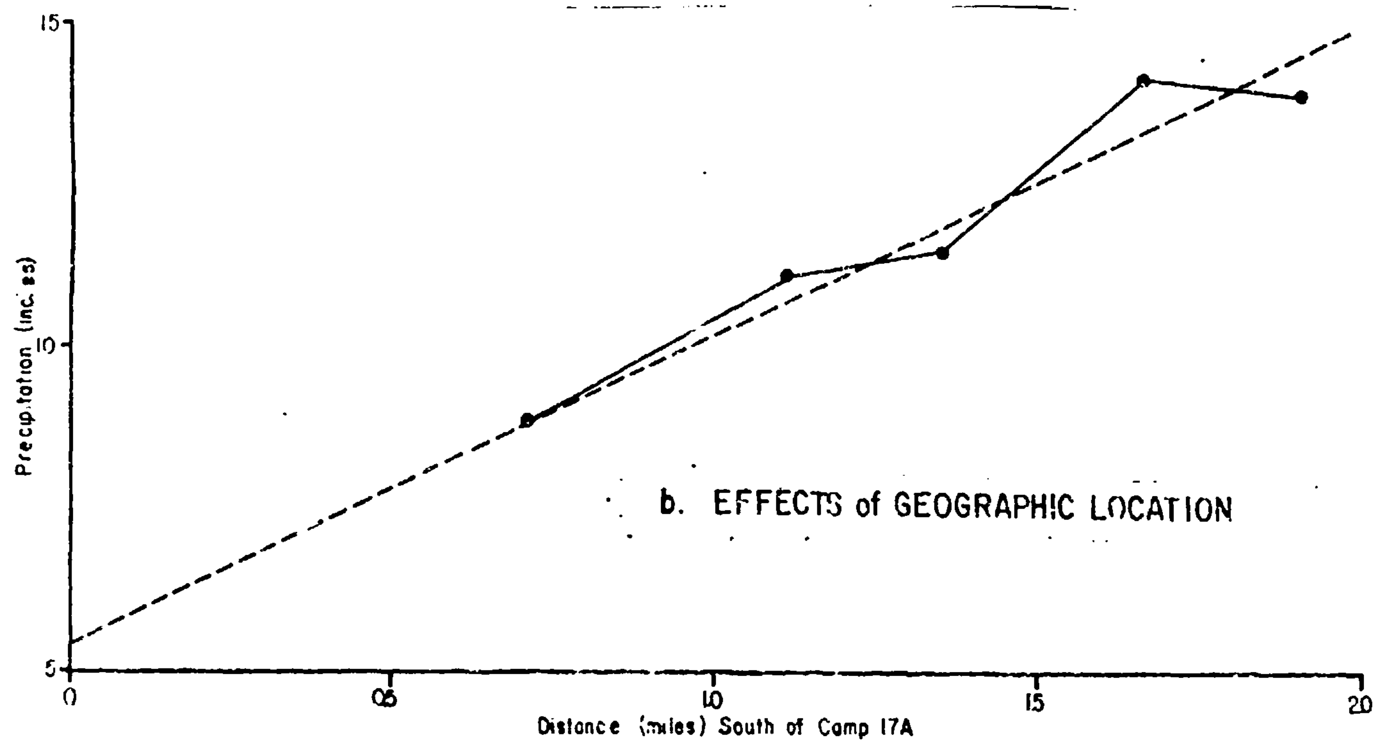
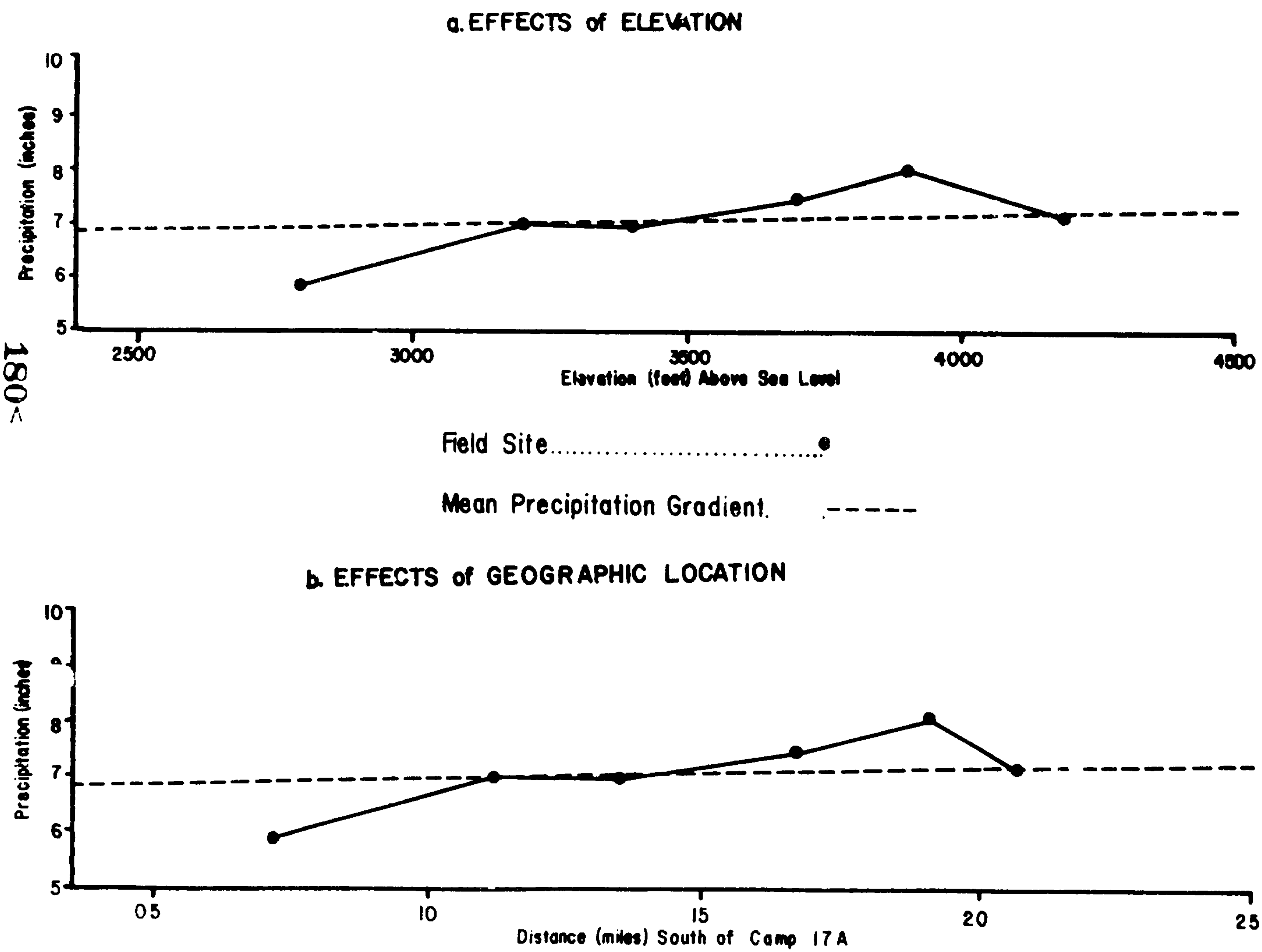


Fig. 40 - Precipitation compared at five stations showing effects of elevation and location for the period 1600 July 27 to 1300 August 8, 1972.

Fig. 41 - Precipitation compared at six stations showing effects of elevation (a) and geographic location (b) for the period 1600 July 27 to 1500 August 4, 1972.



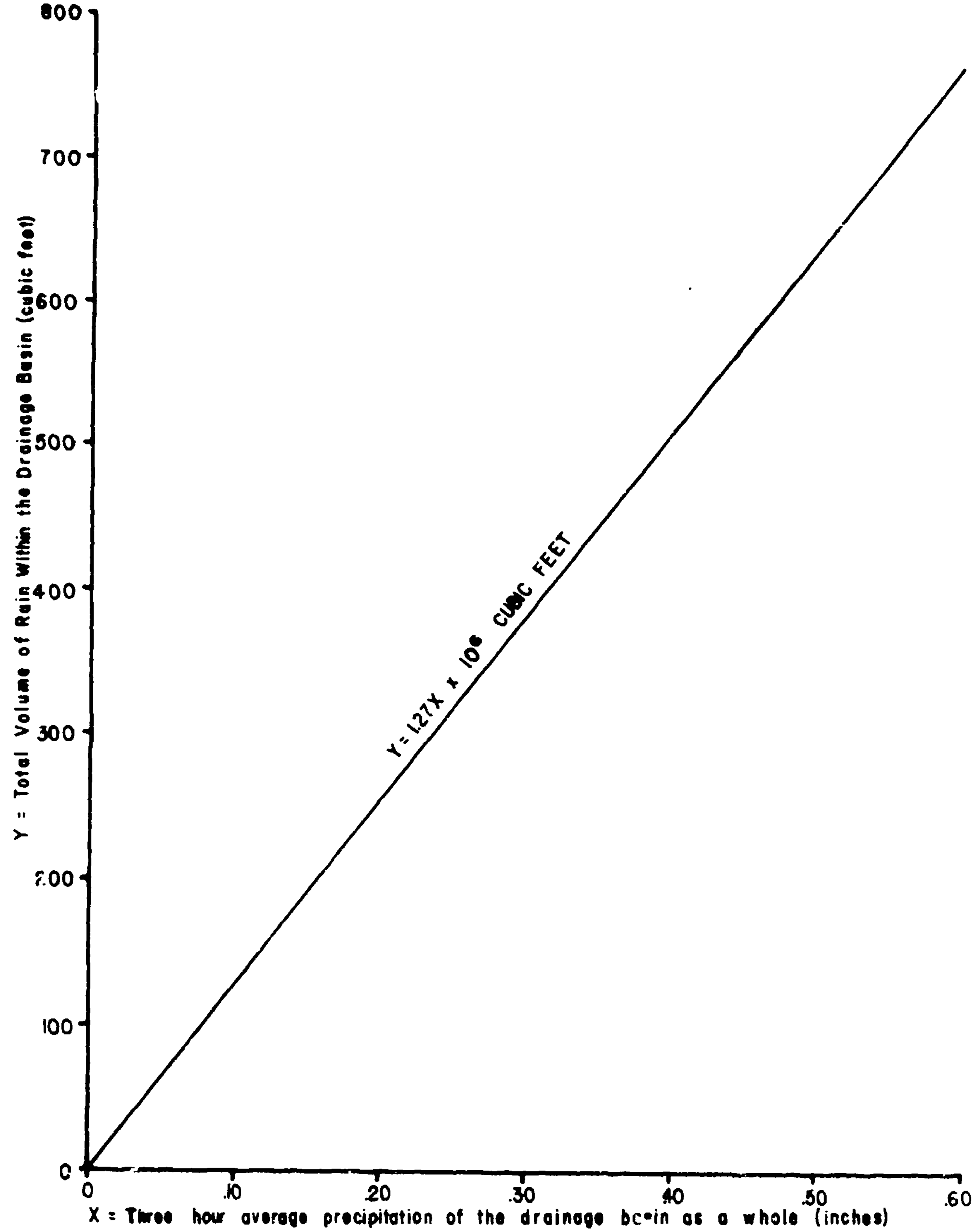


Fig. 42 - Graph of the equation $Y = 1.27X \times 10^6$ cubic feet.

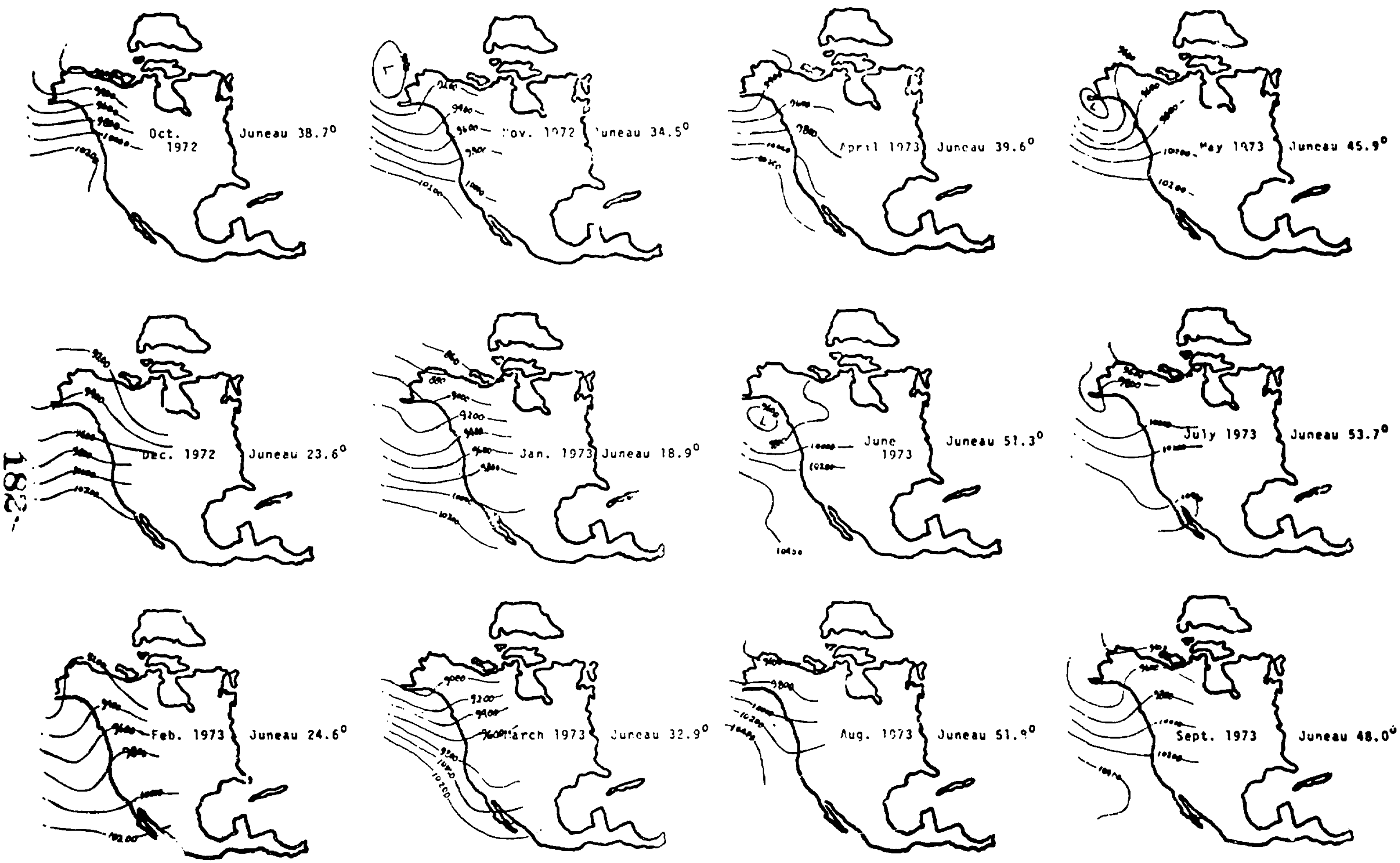
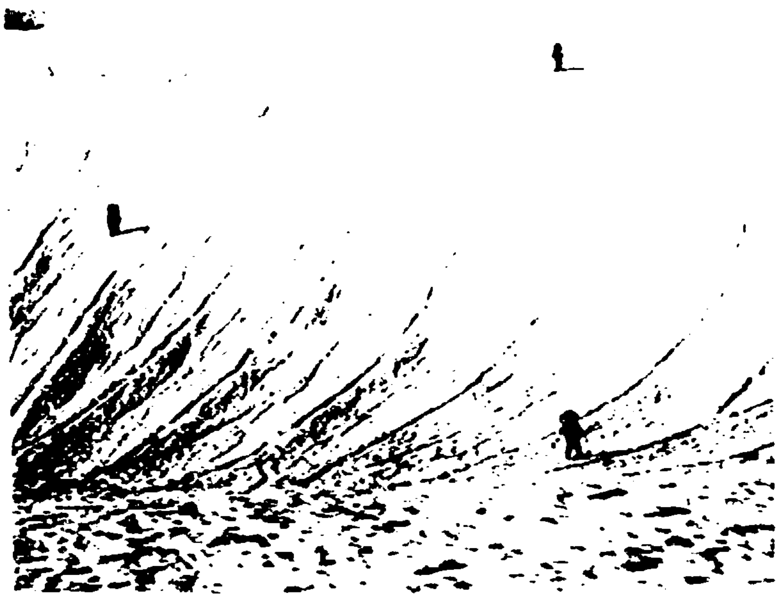


Fig. 43 Mean 700 mb contours over North Pacific Coast, October 1972 to September 1973, based on air flow pattern at about 10,000 feet. Mean monthly temperatures at Juneau, Alaska also shown (USWB data).



Rill and Runnel Pattern on the Lemon Glacier

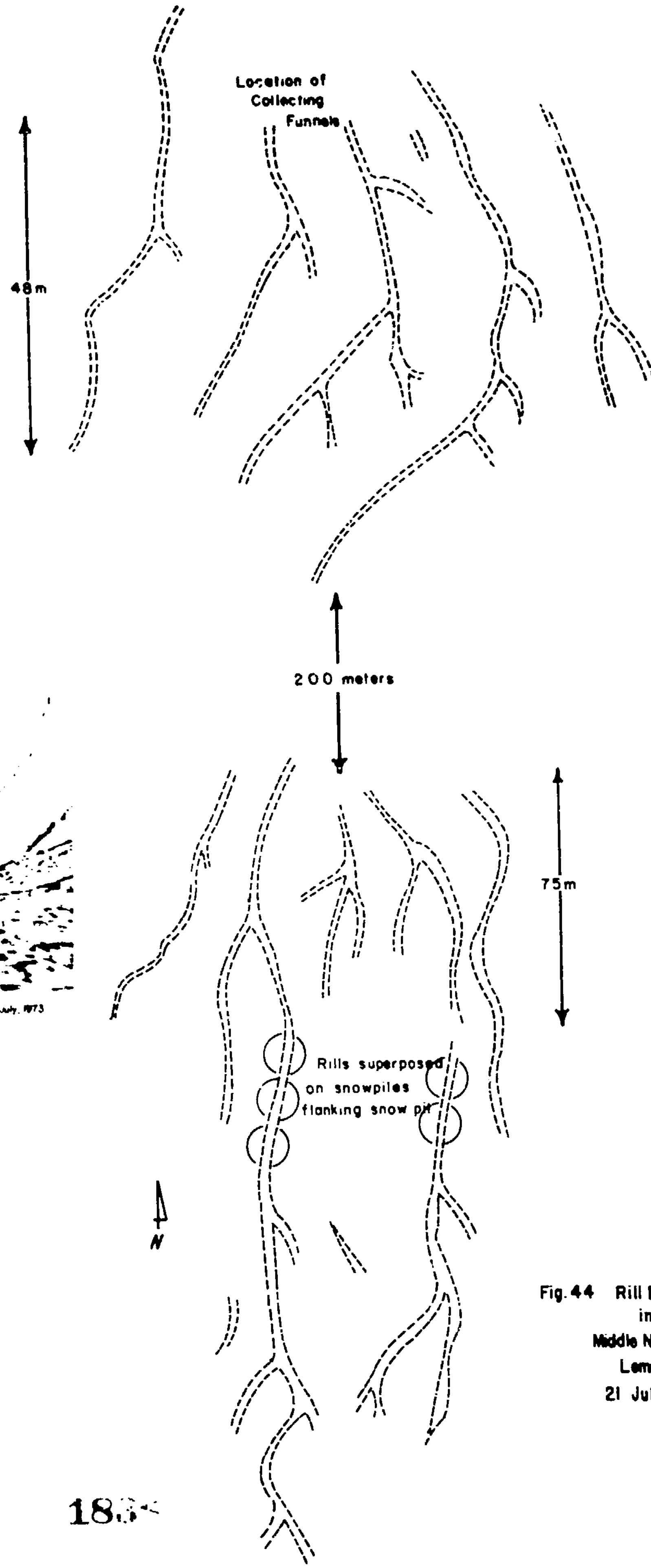
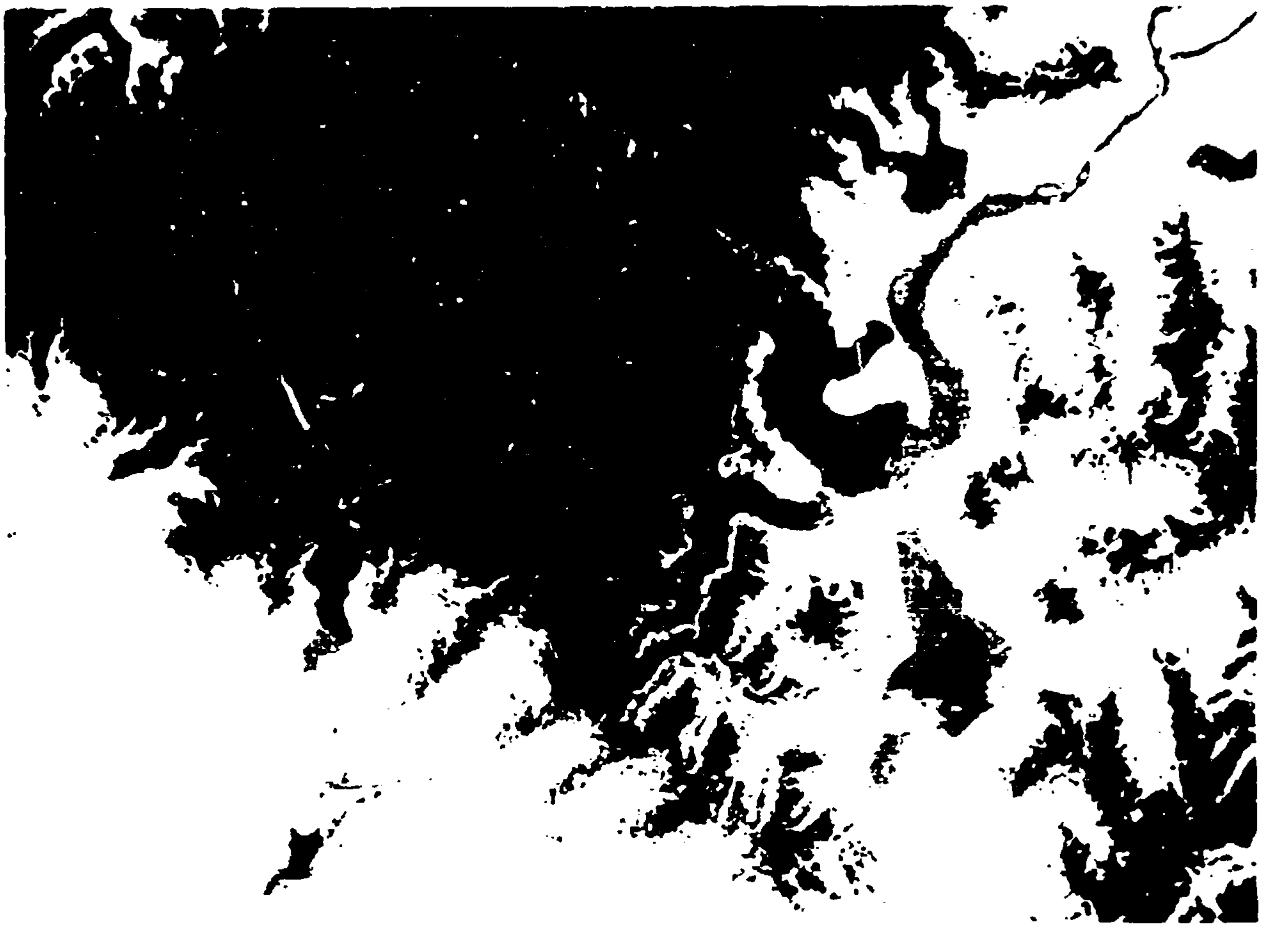
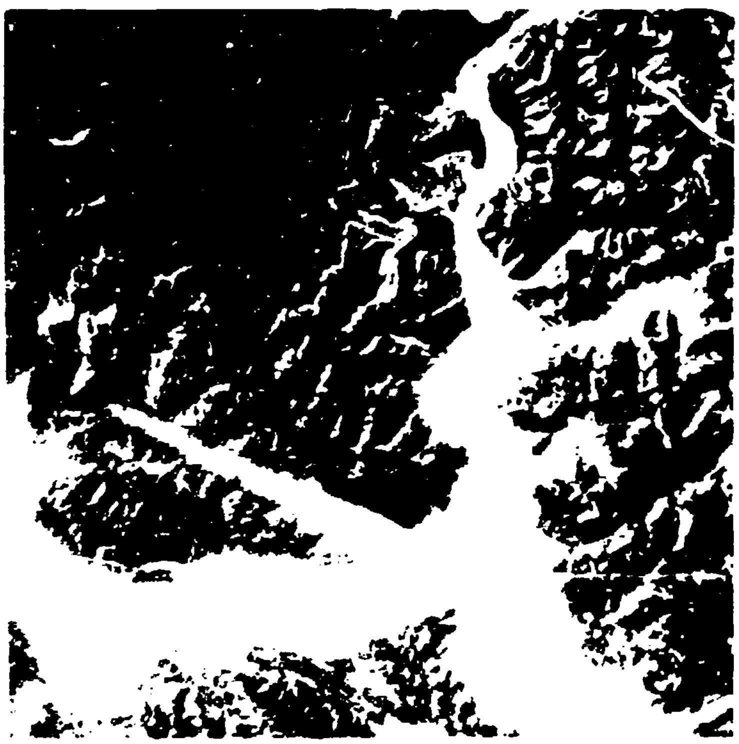


Fig. 44 Rill Patterns in Middle Névé of the Lemon Glacier 21 July 1973



a. Band 5 (lower red imagery) . . . with glacier areas distinctly black. Eagle, Herbert, Mendenhall and Lemon Glacier tongues visible in left and bottom center.



b. Band 7 (near-IR imagery) . . . showing brightness reversal in glacier terminal areas. Norris, Taku and Twin Glaciers visible on right in Taku Inlet sector. Gastineau Channel is straight fiord in lower left center.

Fig. 45 ERTS multi-spectral imagery of southwestern portion of Juneau Icefield, Alaska, on August 11, 1972.

FIG-46 VERTICAL TEST PIT PROFILES ON TAKU AND LLEWELLYN GLACIER CRESTAL NÉVÉS

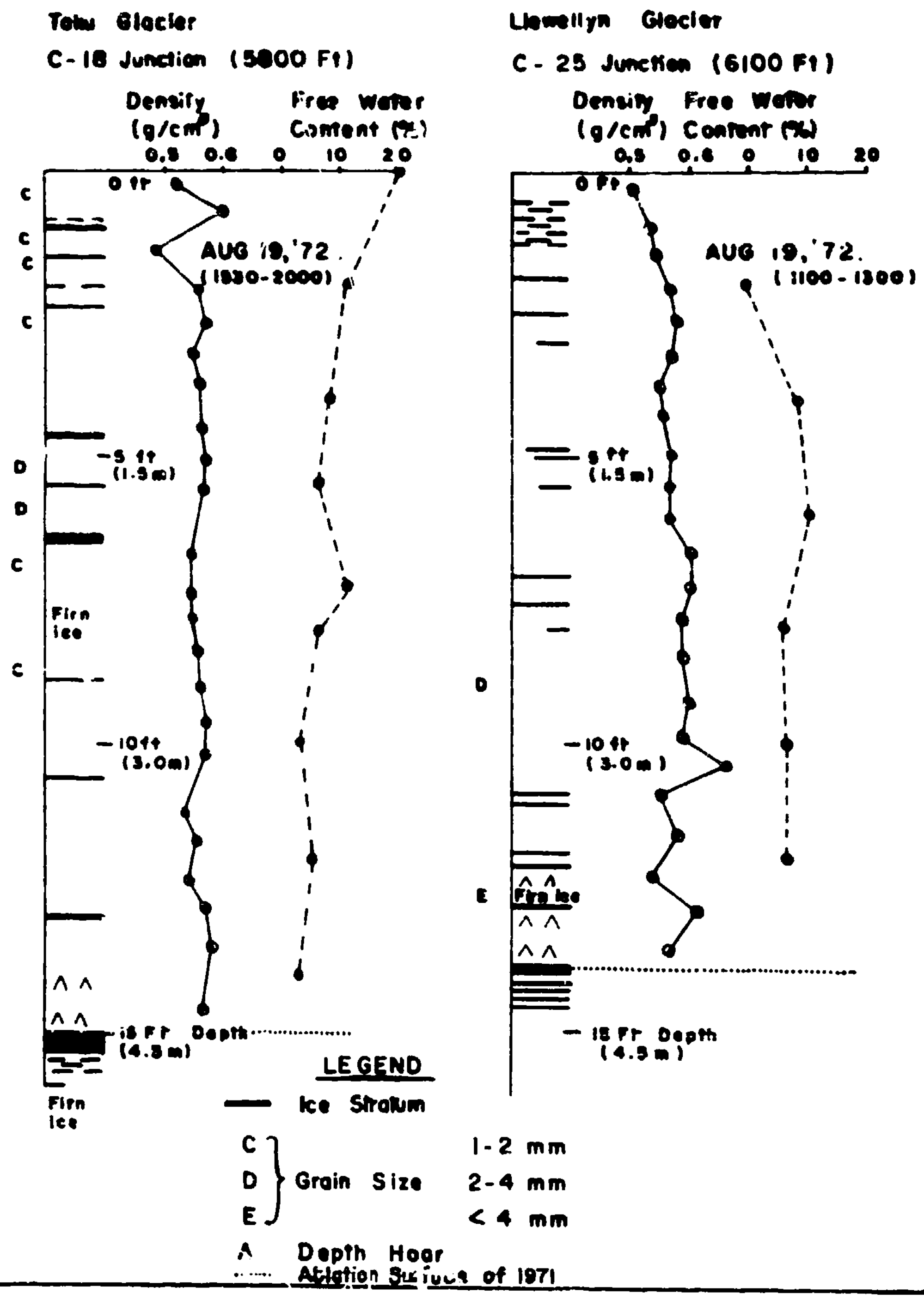


FIG-47 FREE WATER CONTENT OF CATHEDRAL GLACIER FIRN 5850 Ft. (1770 m)

September 7, 1972.

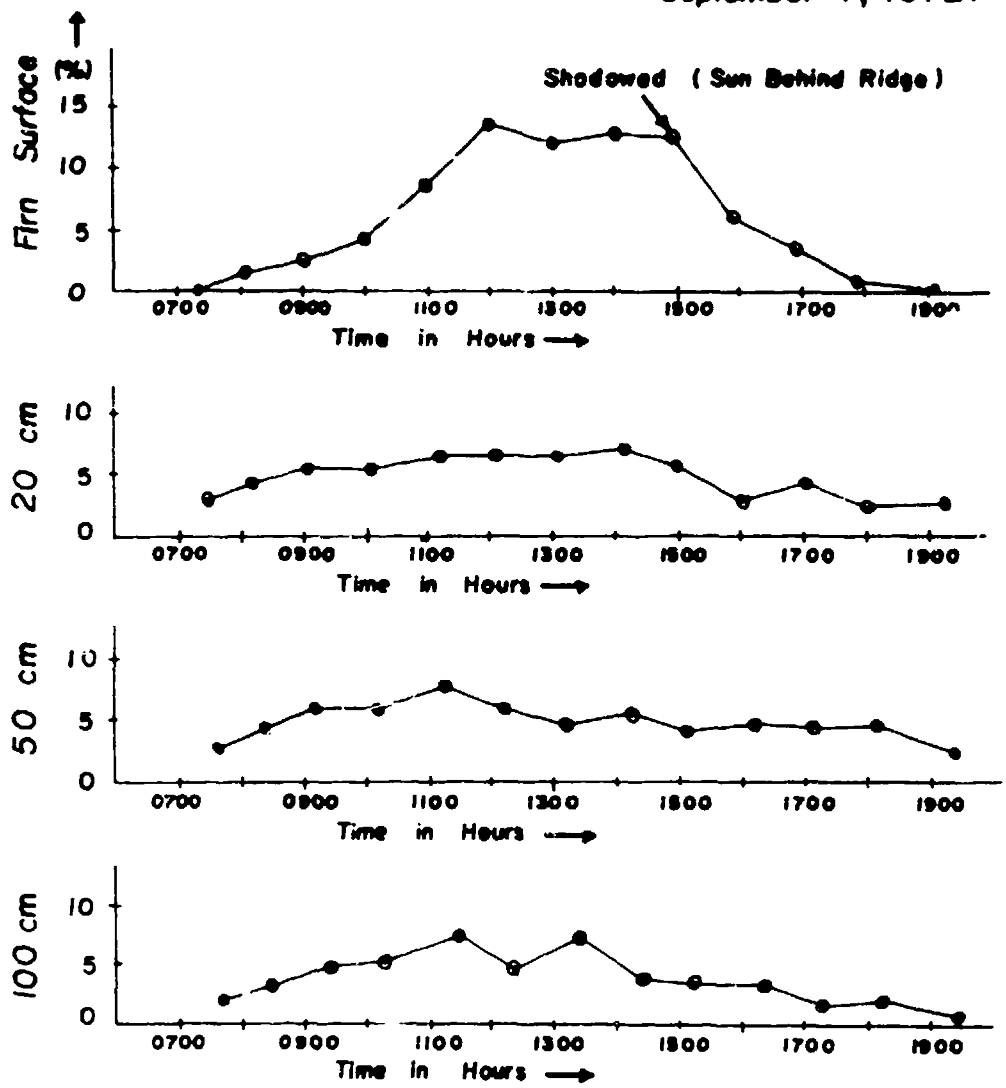


FIG-48 HYDROLOGICAL TREND DURING A DIURNAL CYCLE ON THE CATHEDRAL GLACIER

Sept. 7, 1972

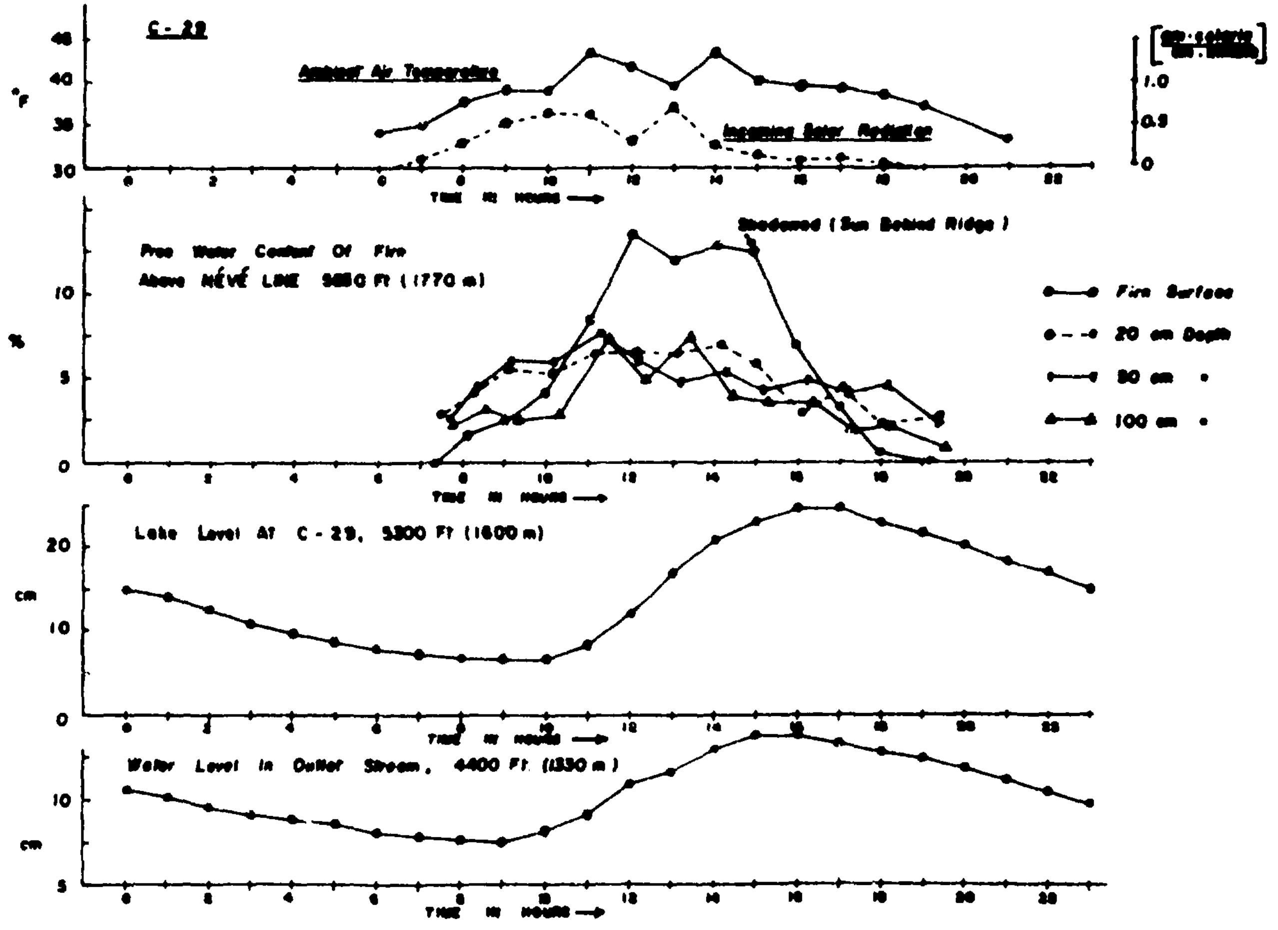
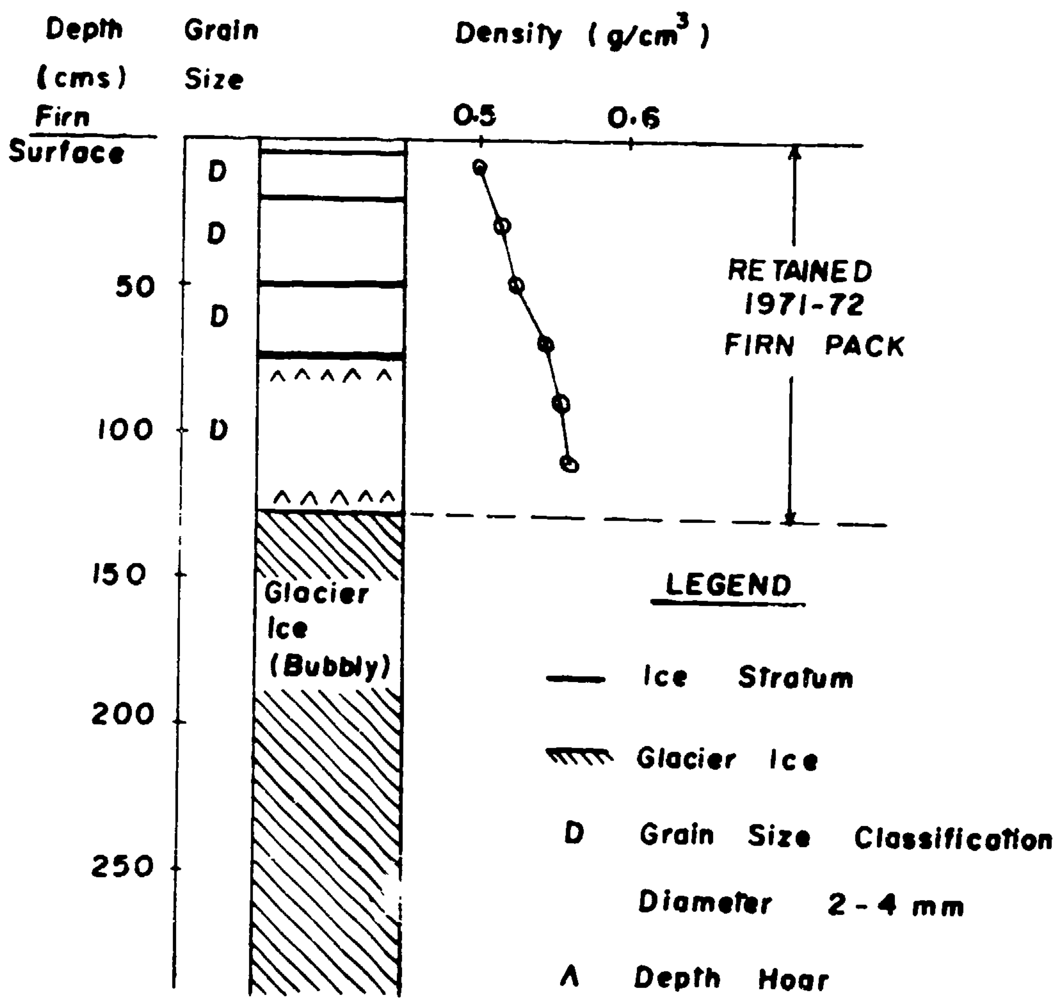


FIG-49 FIRN PIT PROFILE AT CATHEDRAL GLACIER TEST PIT SITE, 5850 Ft. (1770 m) September 7, 1972.



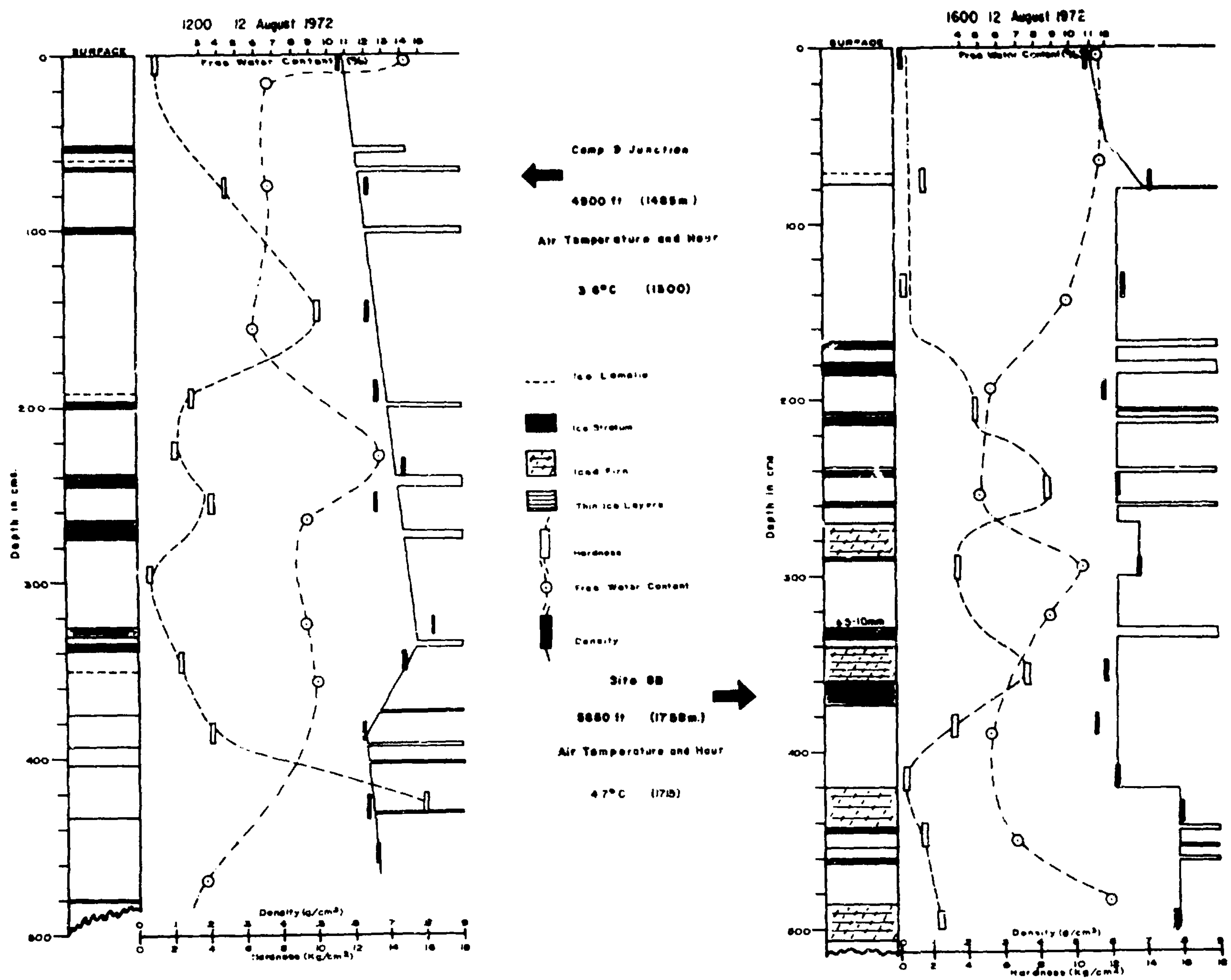


Fig. 50c FIRN STRATIGRAPHY on HIGHER NÉVÉS, JUNEAU ICEFIELD, 1972

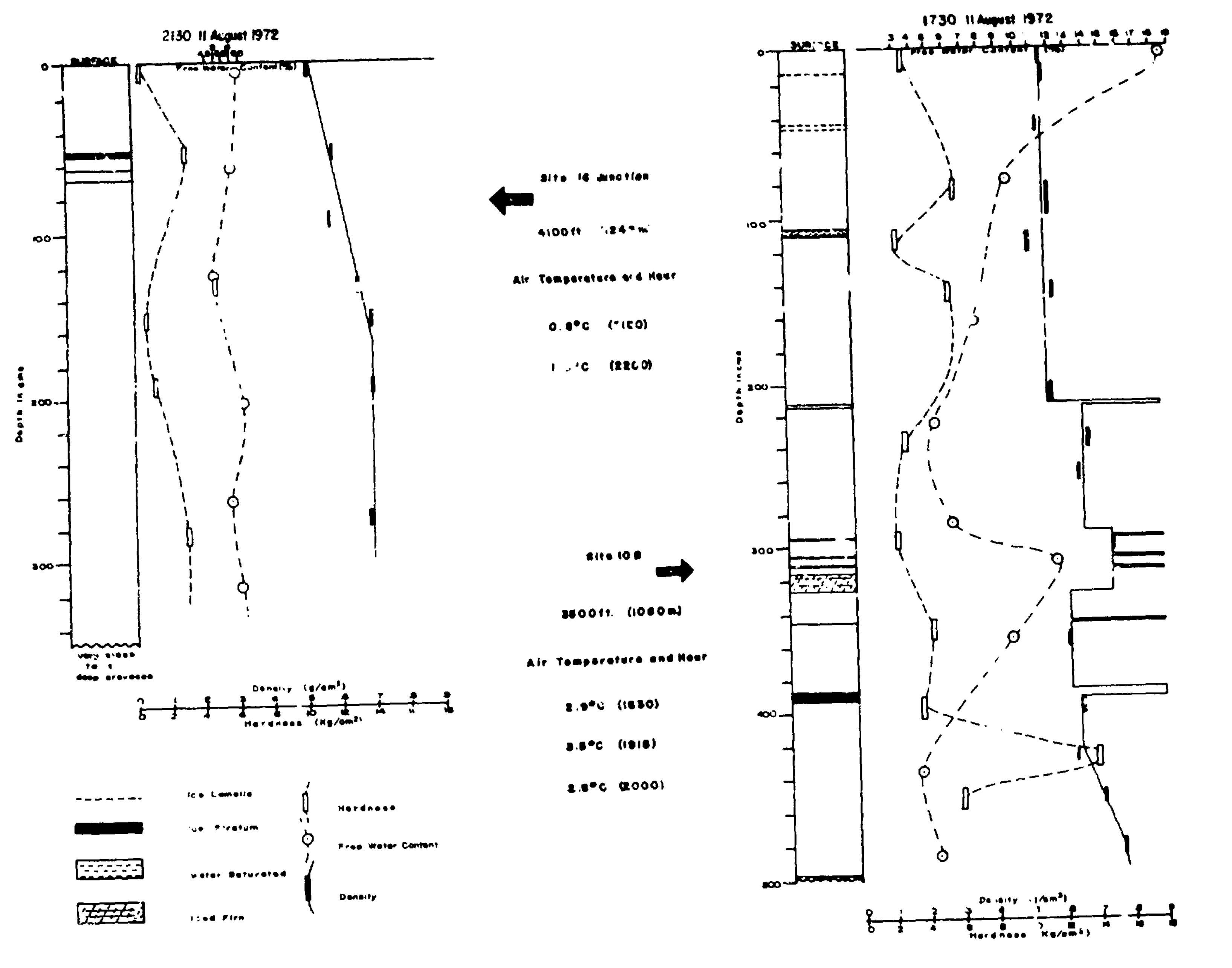


Fig.50b FIRN STRATIGRAPHY on LOWER NÉVÉS, JUNEAU ICEFIELD, 1972

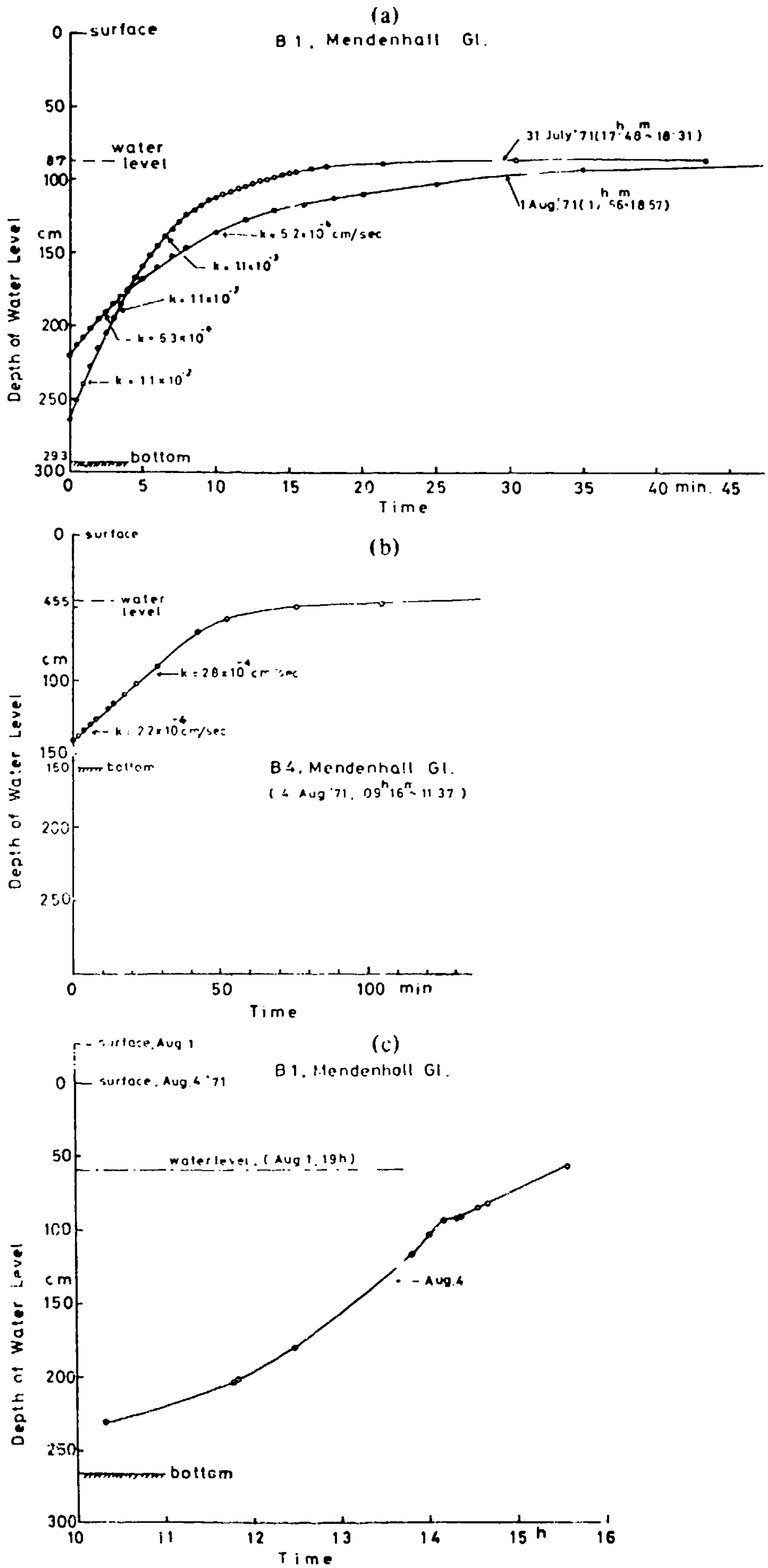


Fig. 51 Variation of water levels with time in bubbly ice near Mendenhall Glacier terminus.

Fig. 52 Fluted walls of melt-water stream on Vaughan Lewis Glacier (photo A. C. Pinchak)

Fig. 53 Oblique aerial view of Vaughan Lewis Glacier Icefall showing wave-bulge zone. Note axial channel of supraglacial stream in wave-ogives. (photo M. M. Miller)

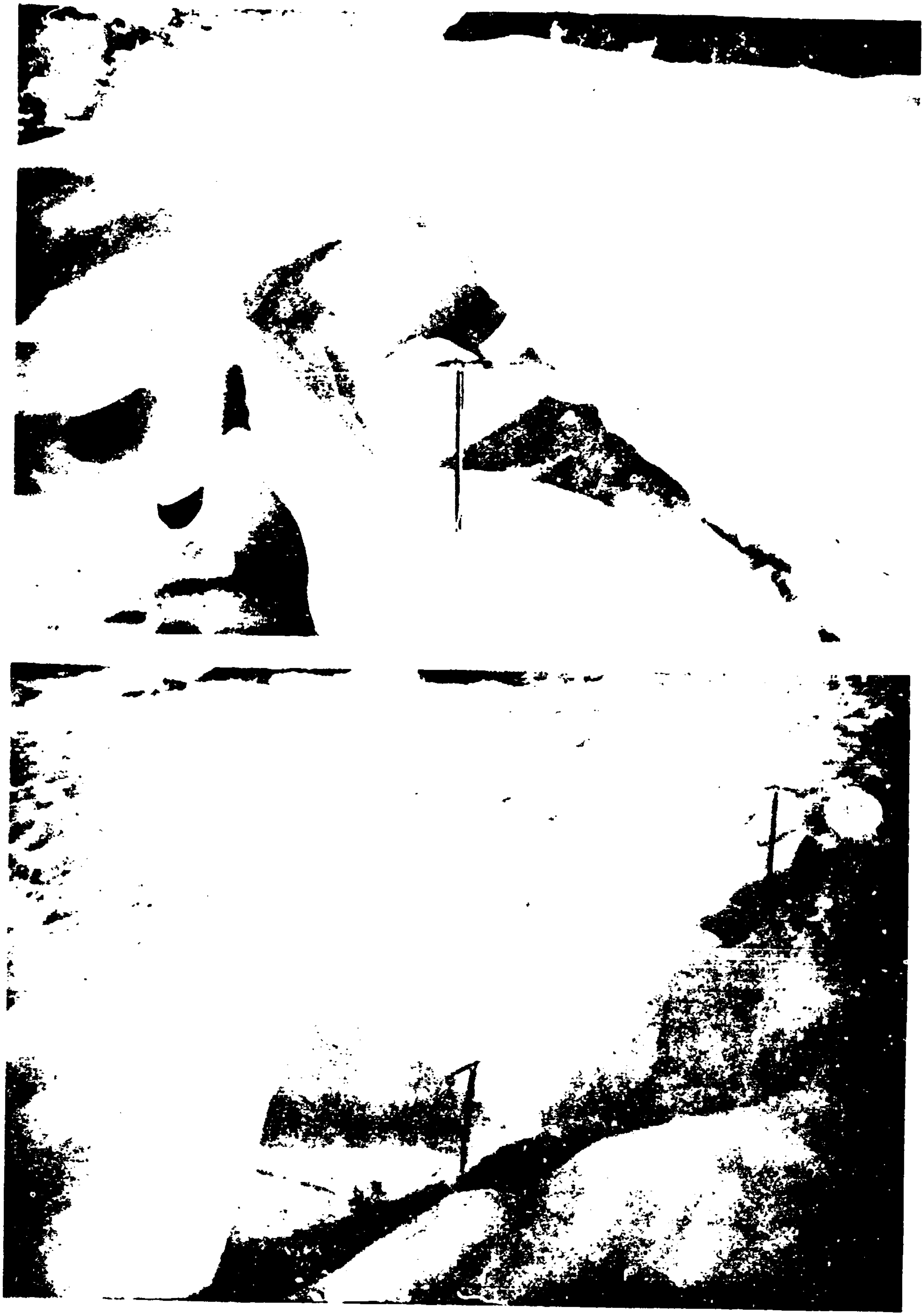
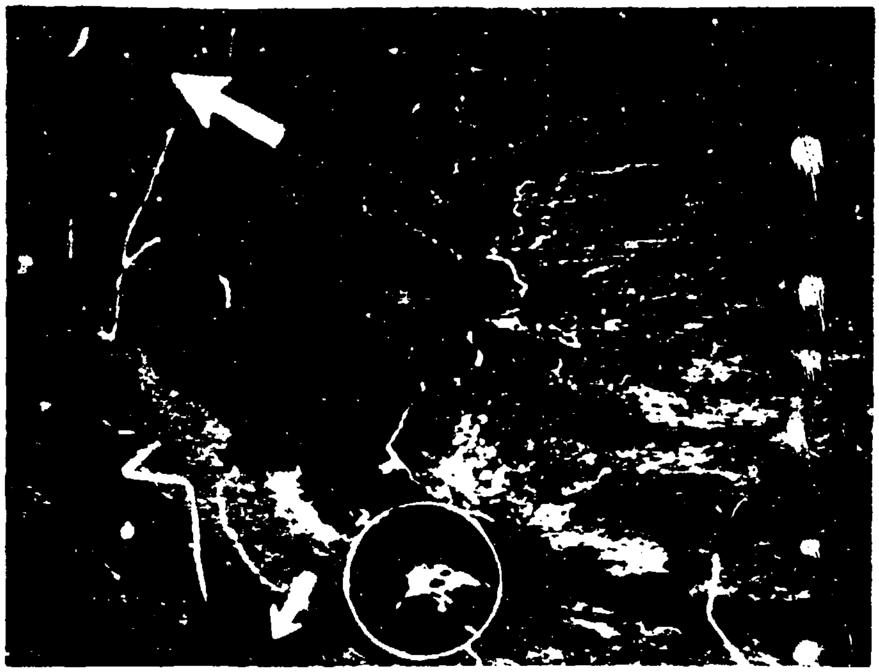
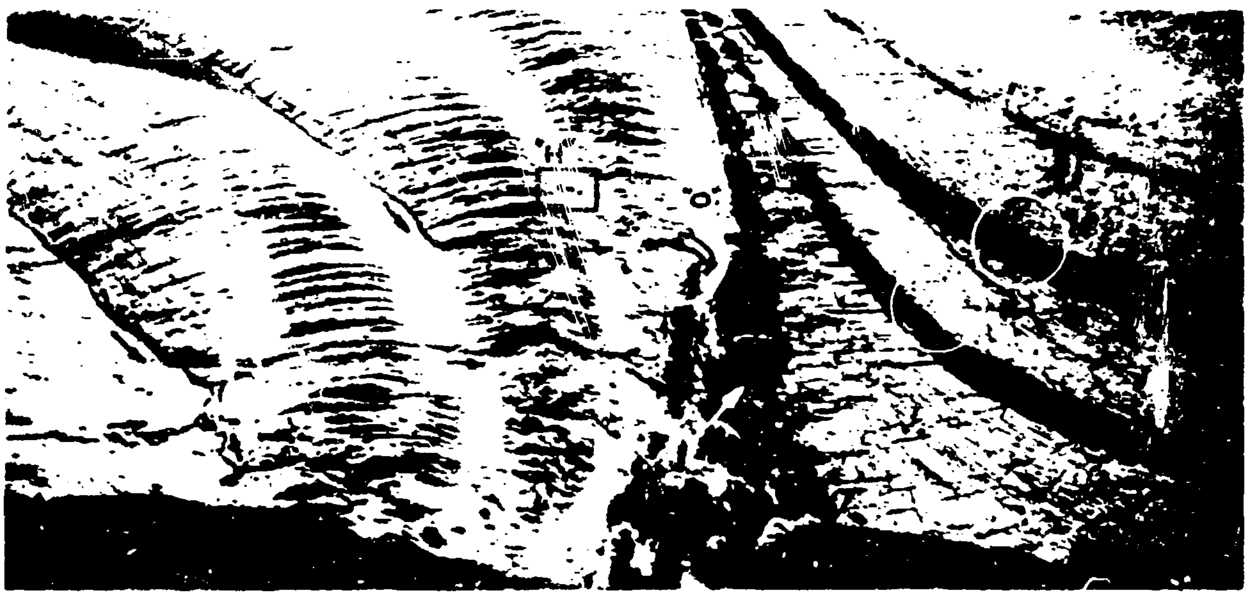


Fig. 54 Two views folded firn-pack in confluence zone of Gilkey and Vaughan Lewis Glacier, August, 1973. (photo A. C. Pinchak)



a. Supraglacial stream system highlighted by specular reflection of setting sun. Large arrow is direction of glacier flow. Small arrow at moulin into which streams flow.



b. Medial moraines and wave-bulge features showing transecting radial crevasse pattern. Circled areas represent sites of detailed study. Scale indicated by human figures appearing as small dots in rectangle H.

Fig. 55 Junction area of Vaughan Lewis Glacier (left) and Gilkey Glacier (right)

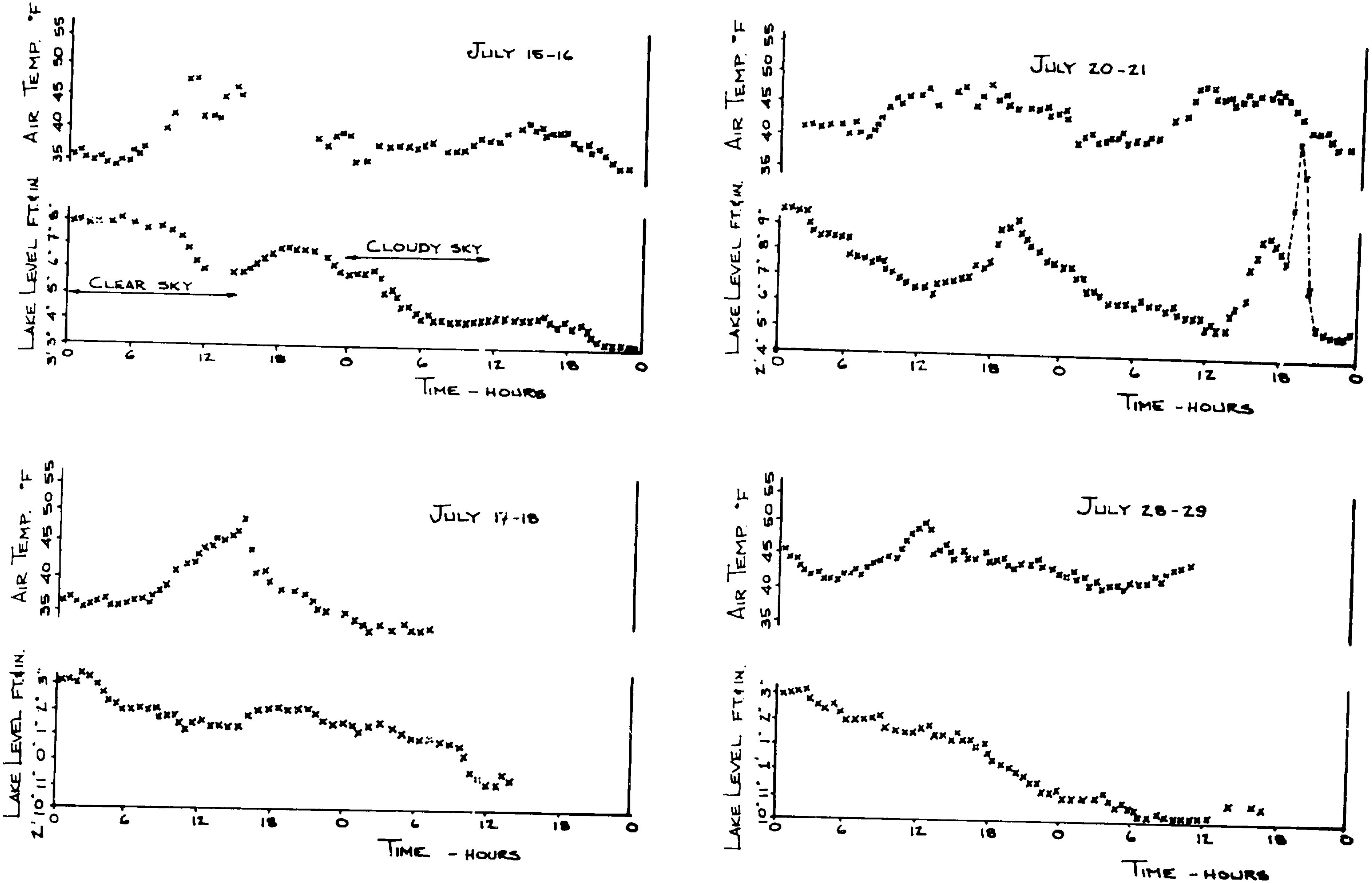


Fig. 56 Water level and ambient air temperature correlations in lakes impounded in wave-bulge depressions of Vaughan Lewis Glacier.

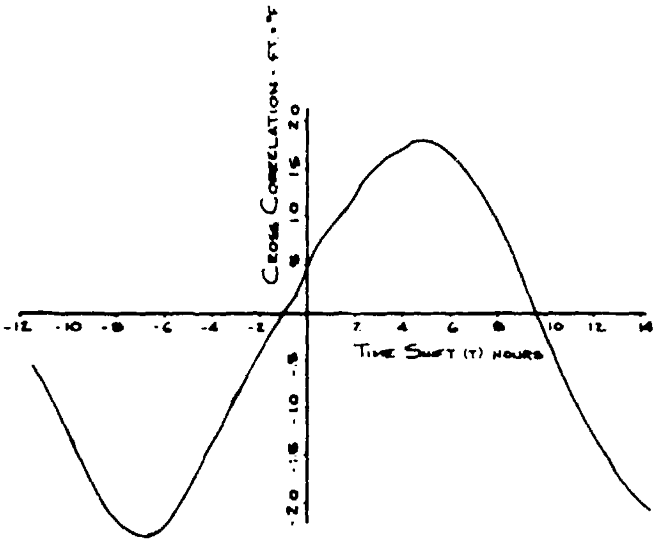


Fig. 57 Cross-correlation of diurnal air temperature changes and stream level variations showing time shift relationship.

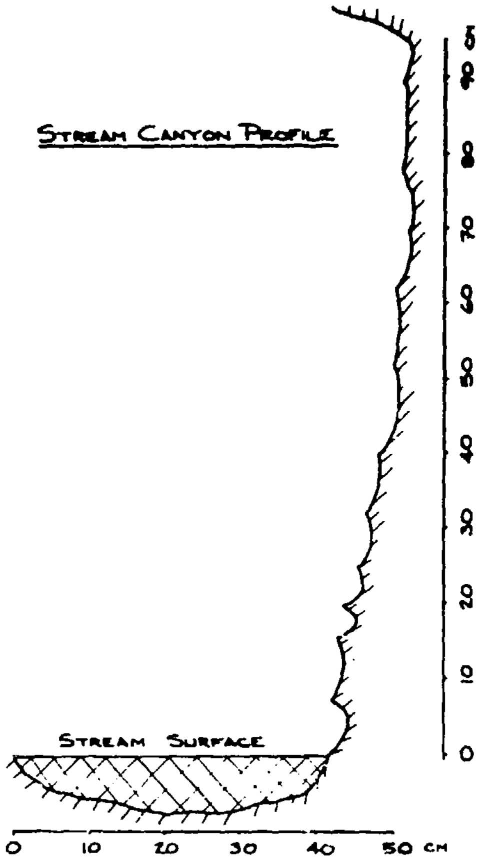


Fig. 59 Side-wall fluting in supraglacial stream channel on Vaughan Lewis Glacier.

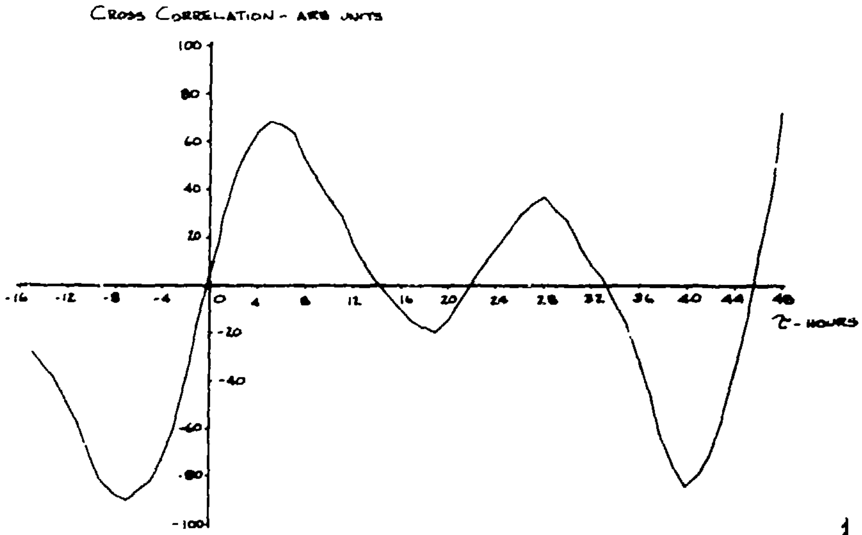


Fig. 58 Cross-correlation curve rising arbitrary units.

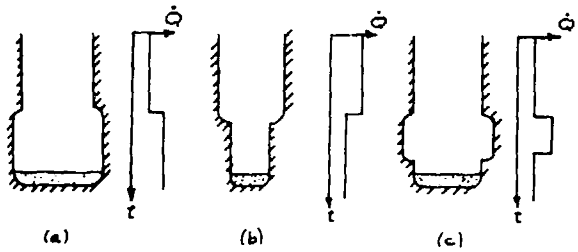


Fig. 60 Schematic representation of side-wall fluting produced by Thermal dissipation from flowing water.

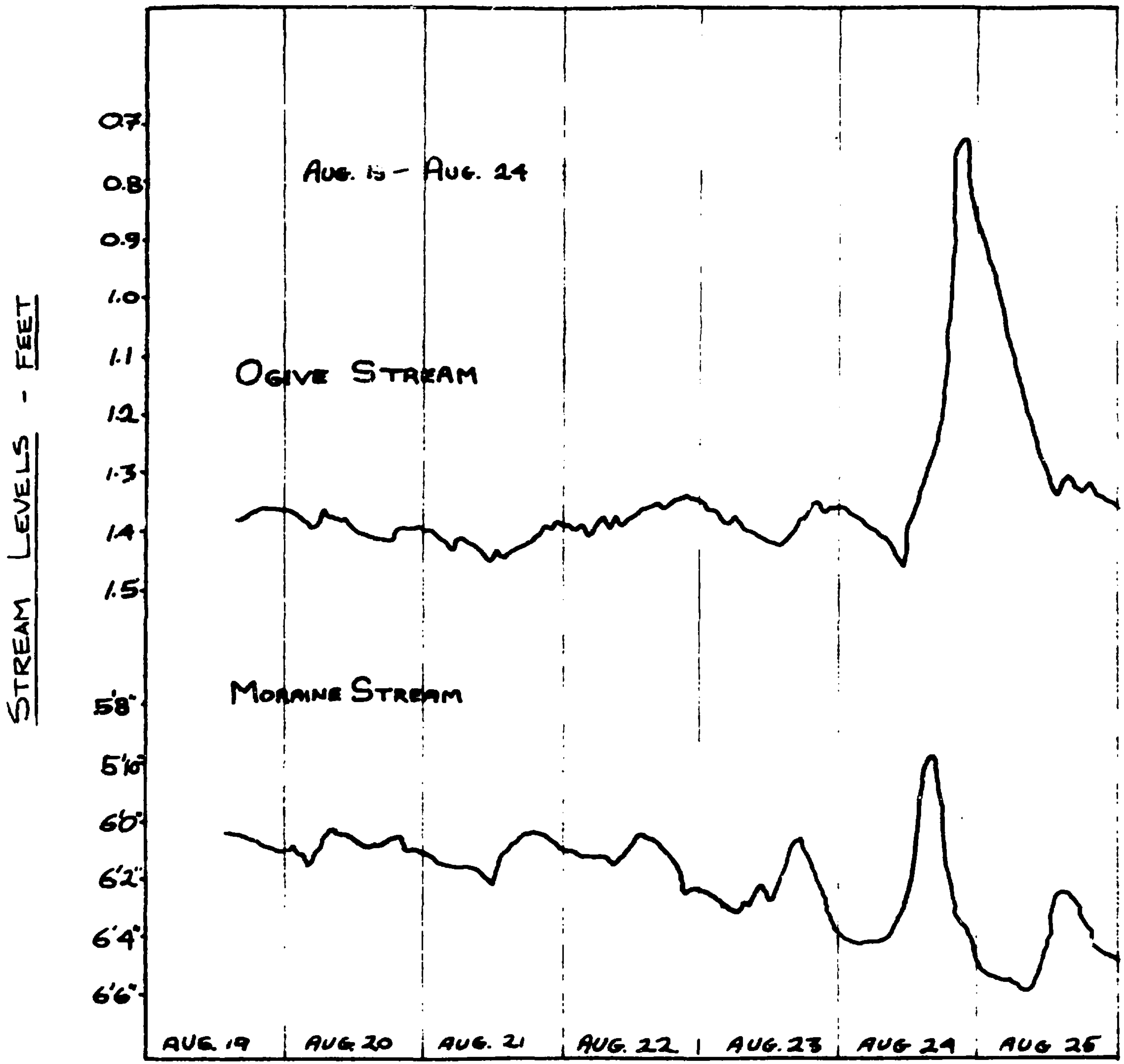


FIG. 61 COMPARISON OF STREAM LEVELS IN TWO SECTORS OF THE VAUGHAN LEWIS GLACIER

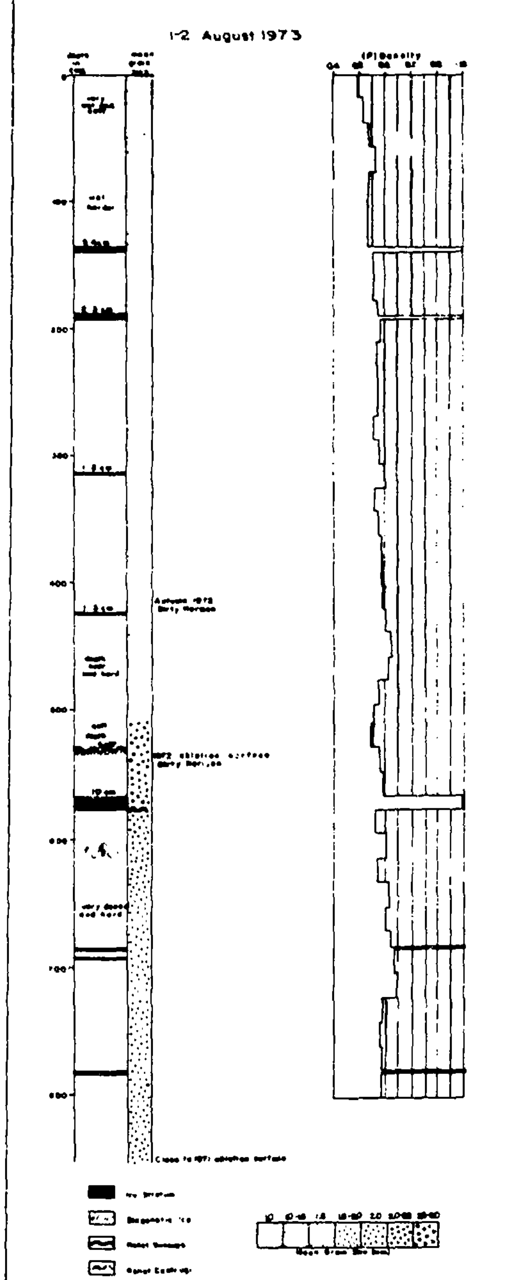


Fig. 62a FIRN STRATIGRAPHY on HIGHER NÉVÉS
Juneau Icefield, 1973
Site 8B

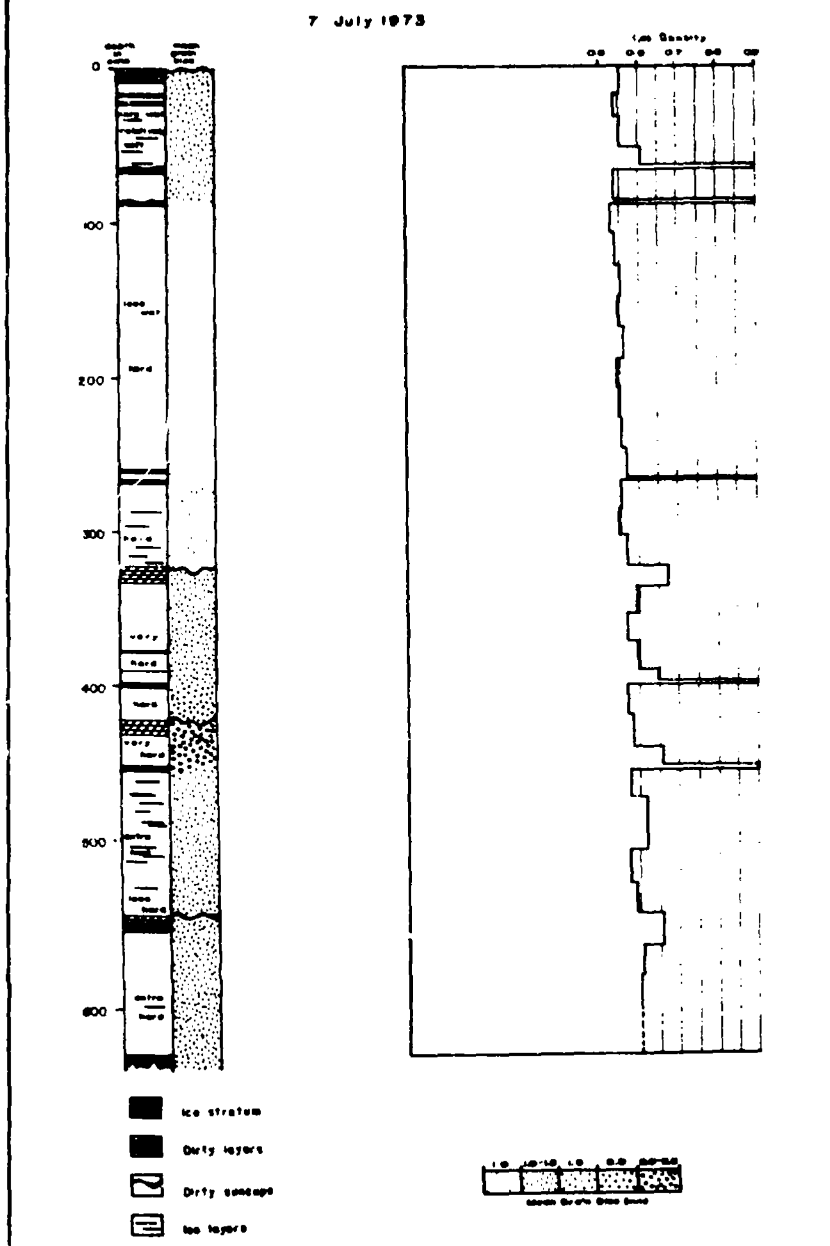


Fig 62b Firn Stratigraphy on Lower Névés
Juneau Icefield, 1973
Site 10B elev 3600 ft.
(1090m)

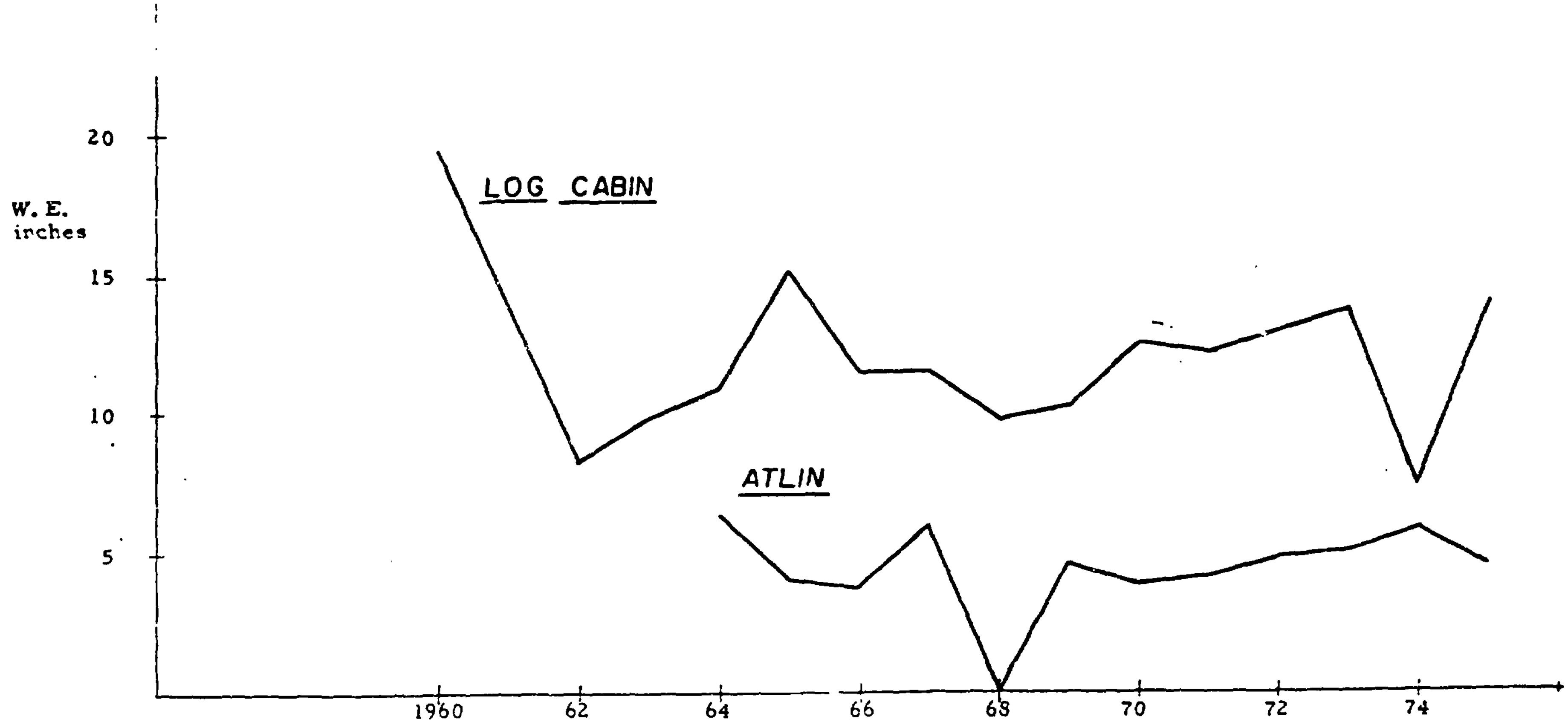
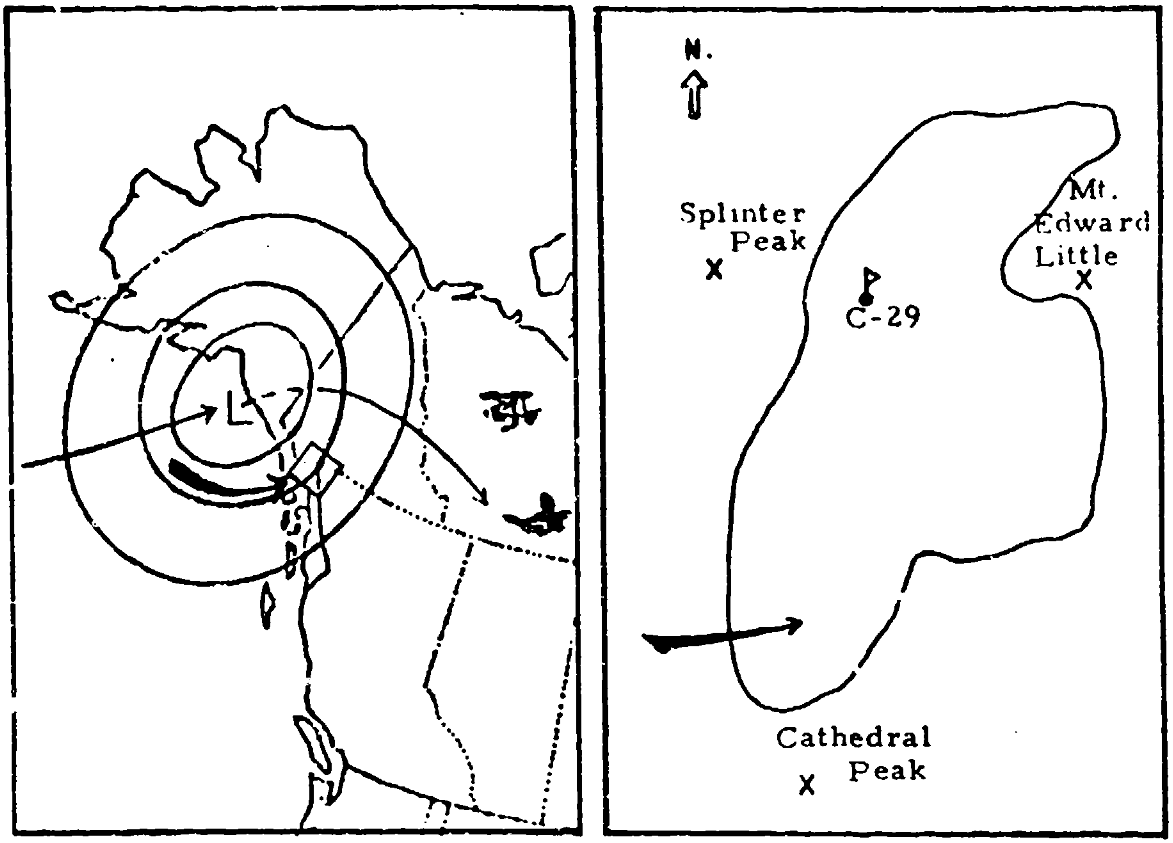
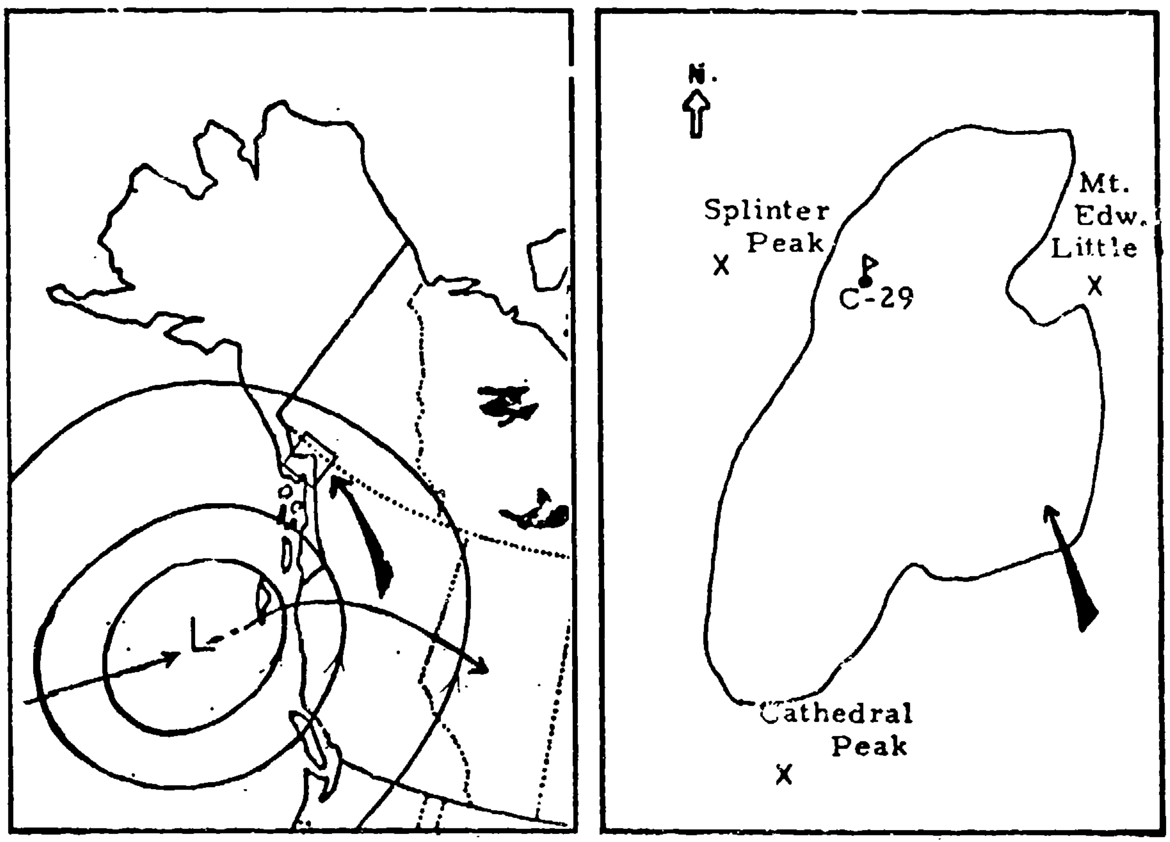


Fig. 63 April 1st Snow depths, Atlin and Log Cabin ; Water Equiv. inches



(a) Cyclonic track passes northward, bringing southwest storm winds



(b) Cyclonic track passes southward, bringing southeast storm winds

FIG. 64 REPRESENTATION OF STORM WIND SHIFT ON CATHEDRAL GLACIER AND JUNEAU ICEFIELD RESULTING FROM CHANGE IN MEAN POSITION OF CYCLONIC CENTERS (SUB-POLAR LOWS) IN THE GULF OF ALASKA

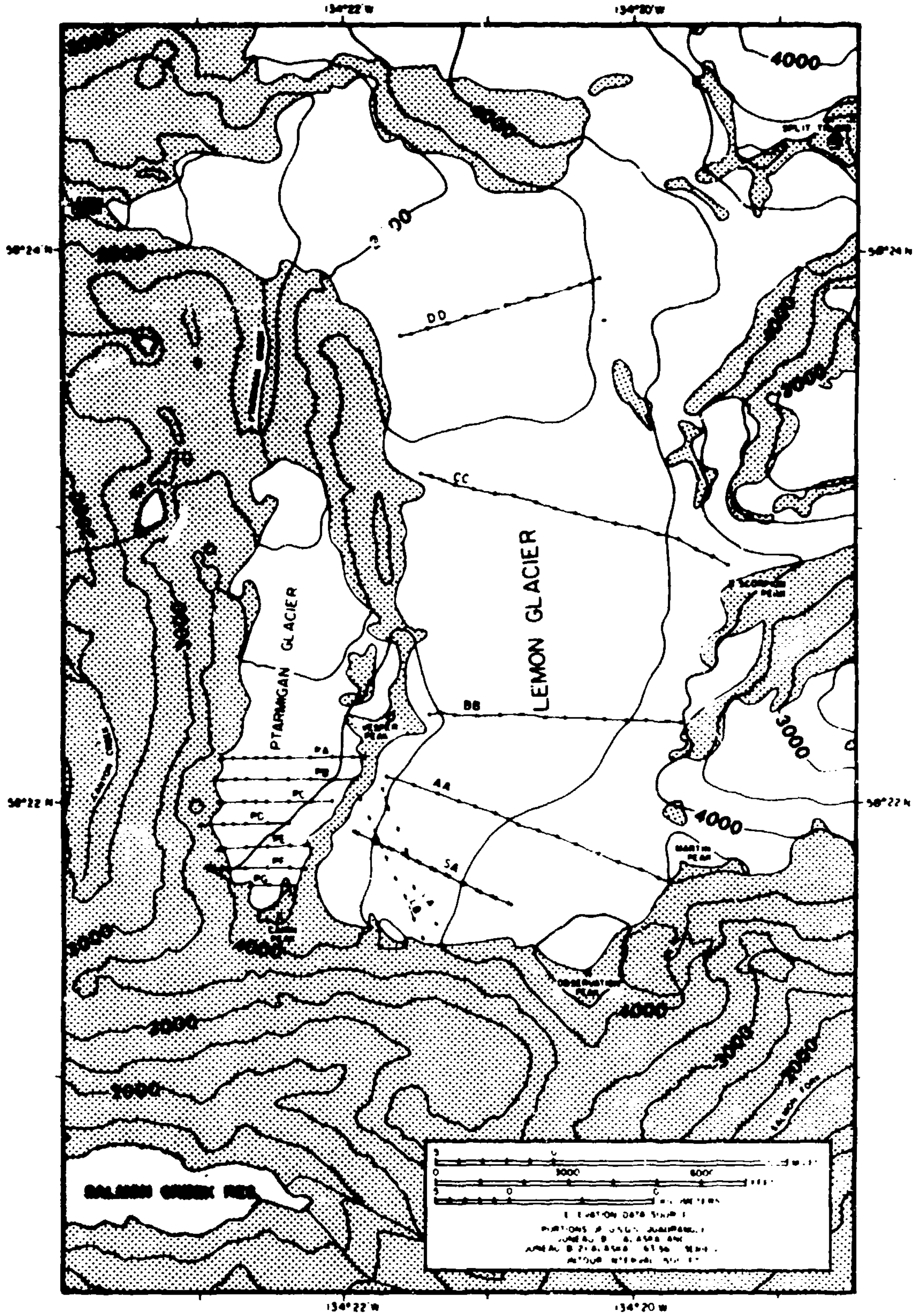


FIG. 65 LOCATION OF GEOPHYSICAL CONTROL
AND GRAVITY-SEISMIC TRAVERSES
ON LEMON AND PTARMIGAN GLACIERS

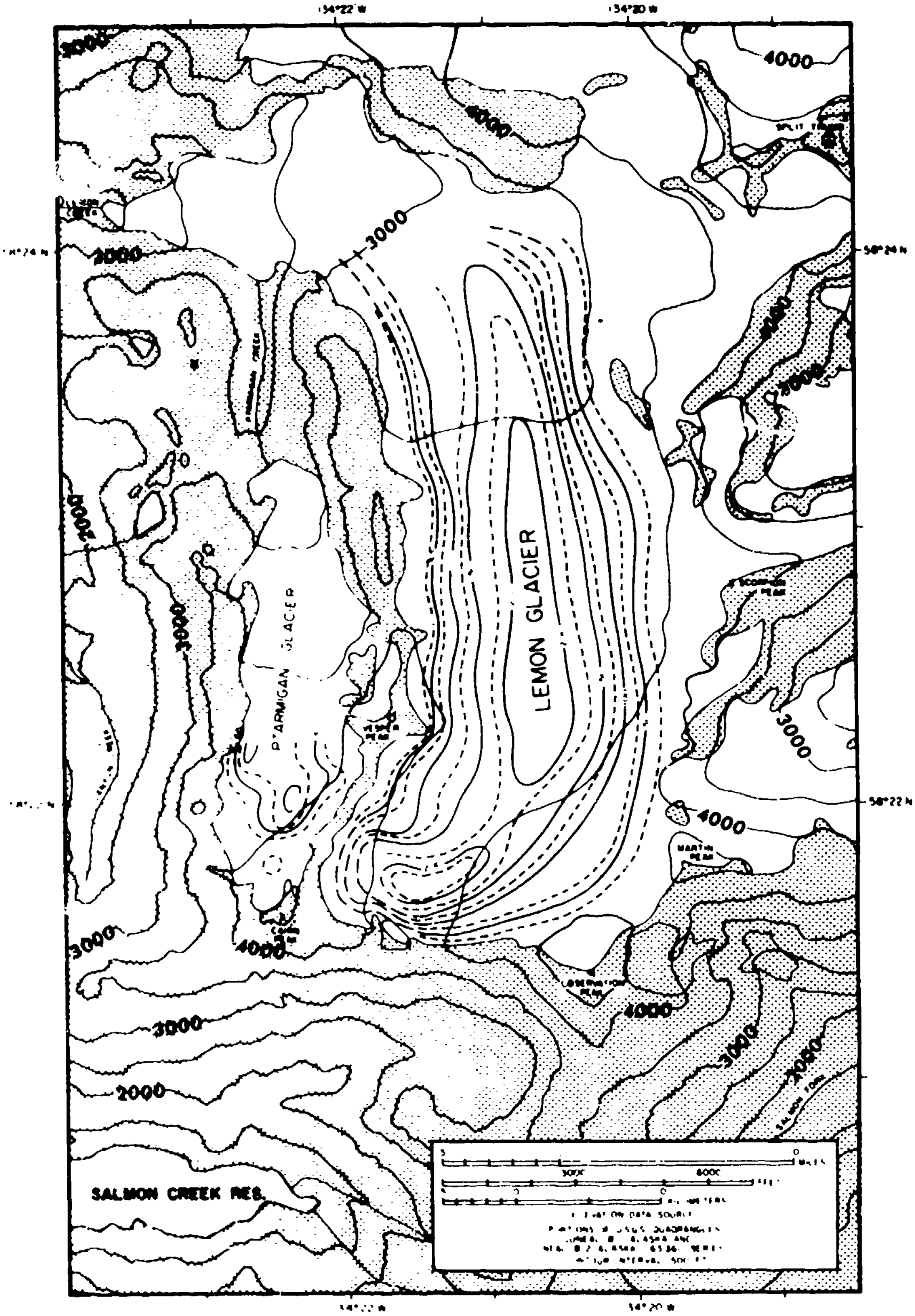


FIG.66 ICE THICKNESS AND SUBGLACIAL TOPOGRAPHY LEMON-PTARMIGAN GLACIER SYSTEM

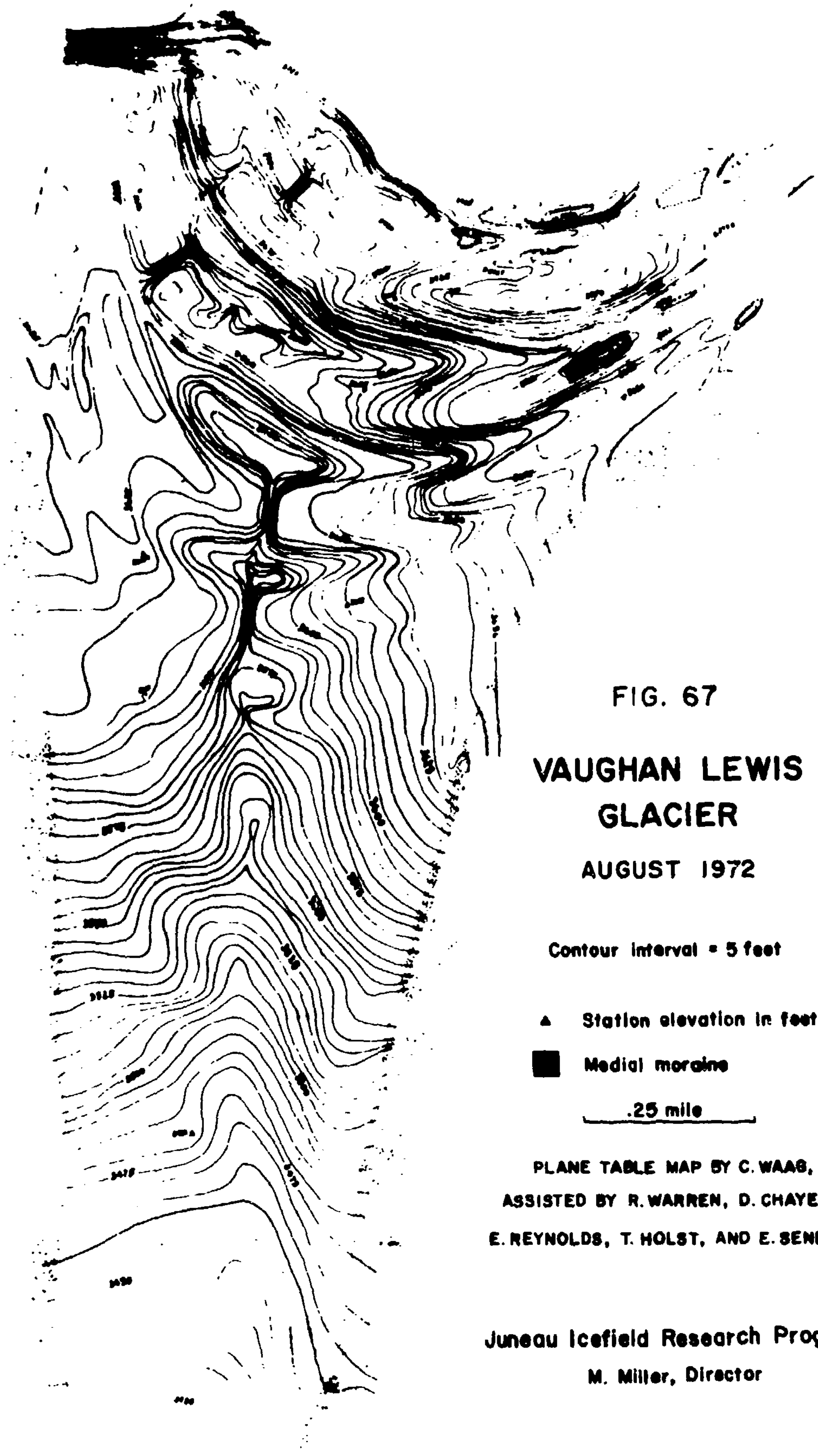


FIG. 67

VAUGHAN LEWIS
GLACIER

AUGUST 1972

Contour interval = 5 feet

▲ Station elevation in feet

■ Medial moraine

.25 mile

PLANE TABLE MAP BY C. WAAG,
ASSISTED BY R. WARREN, D. CHAYES,
E. REYNOLDS, T. HOLST, AND E. SENEAR

Juneau Icefield Research Program
M. Miller, Director

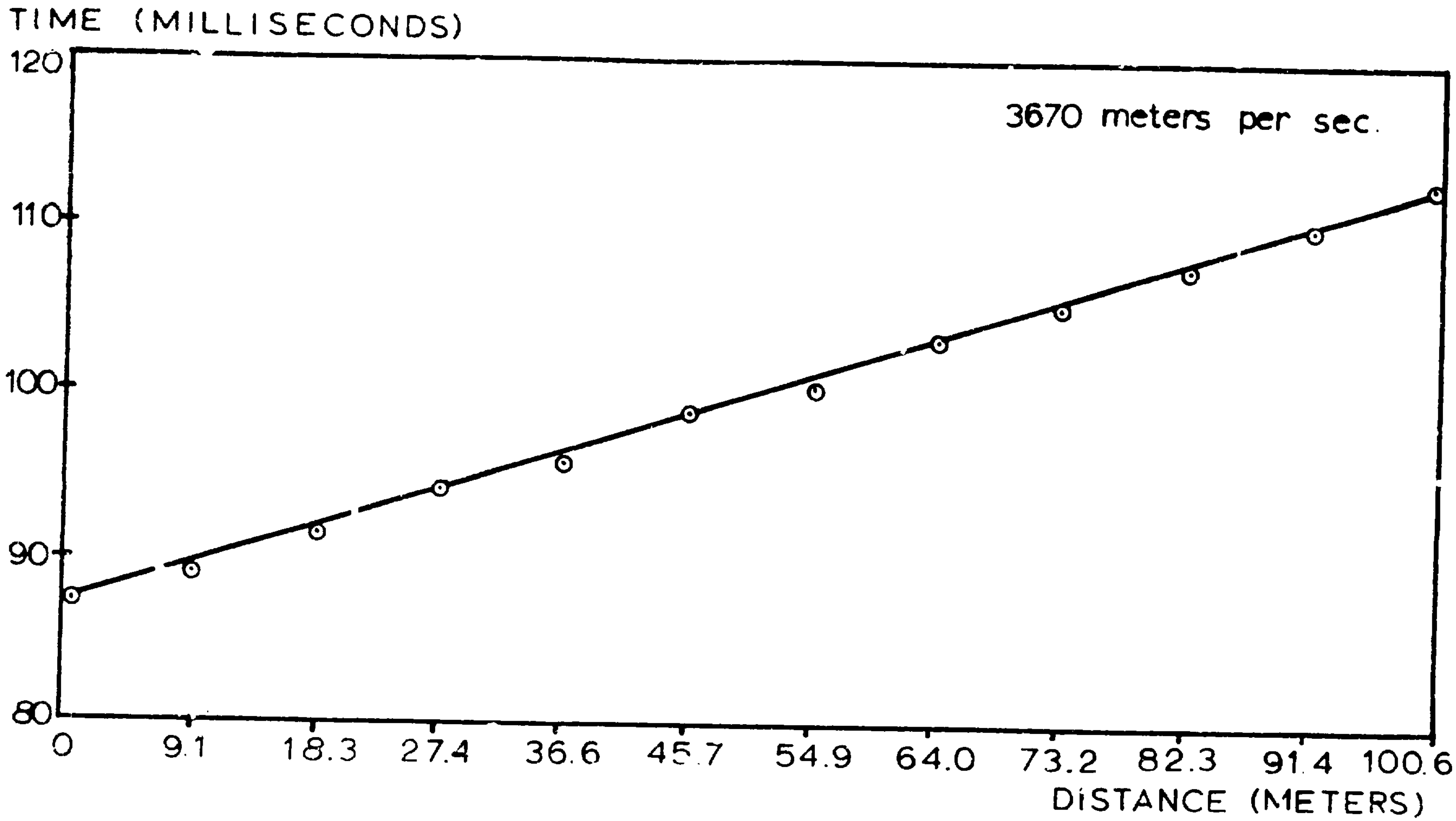


FIG. 68 REFRACTION PROFILE VELOCITY PLOT

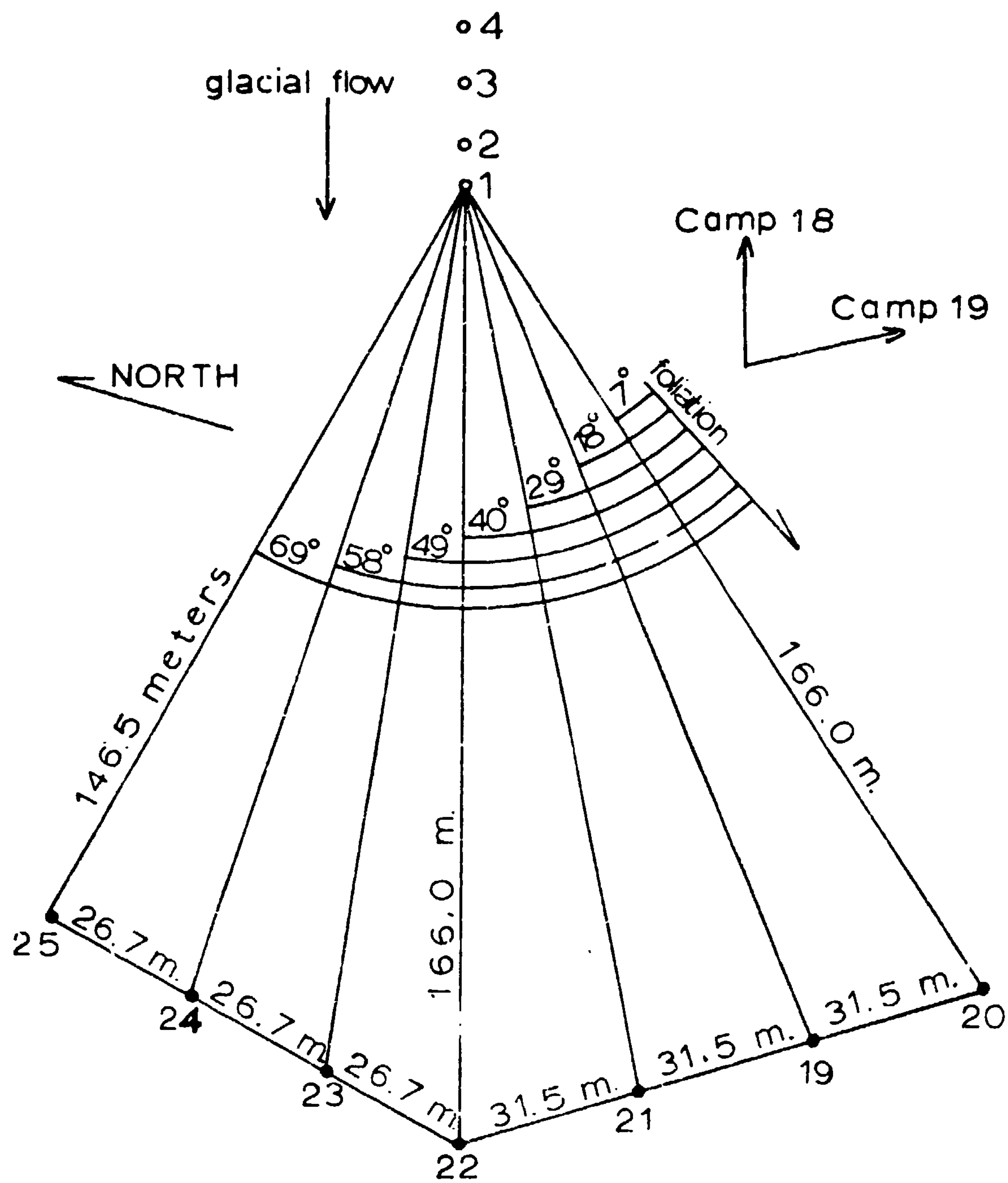
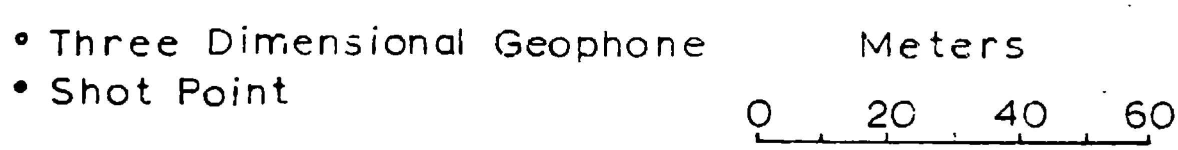


FIG. 69-SEISMIC ARRAY on VAUGHAN LEWIS GLACIER



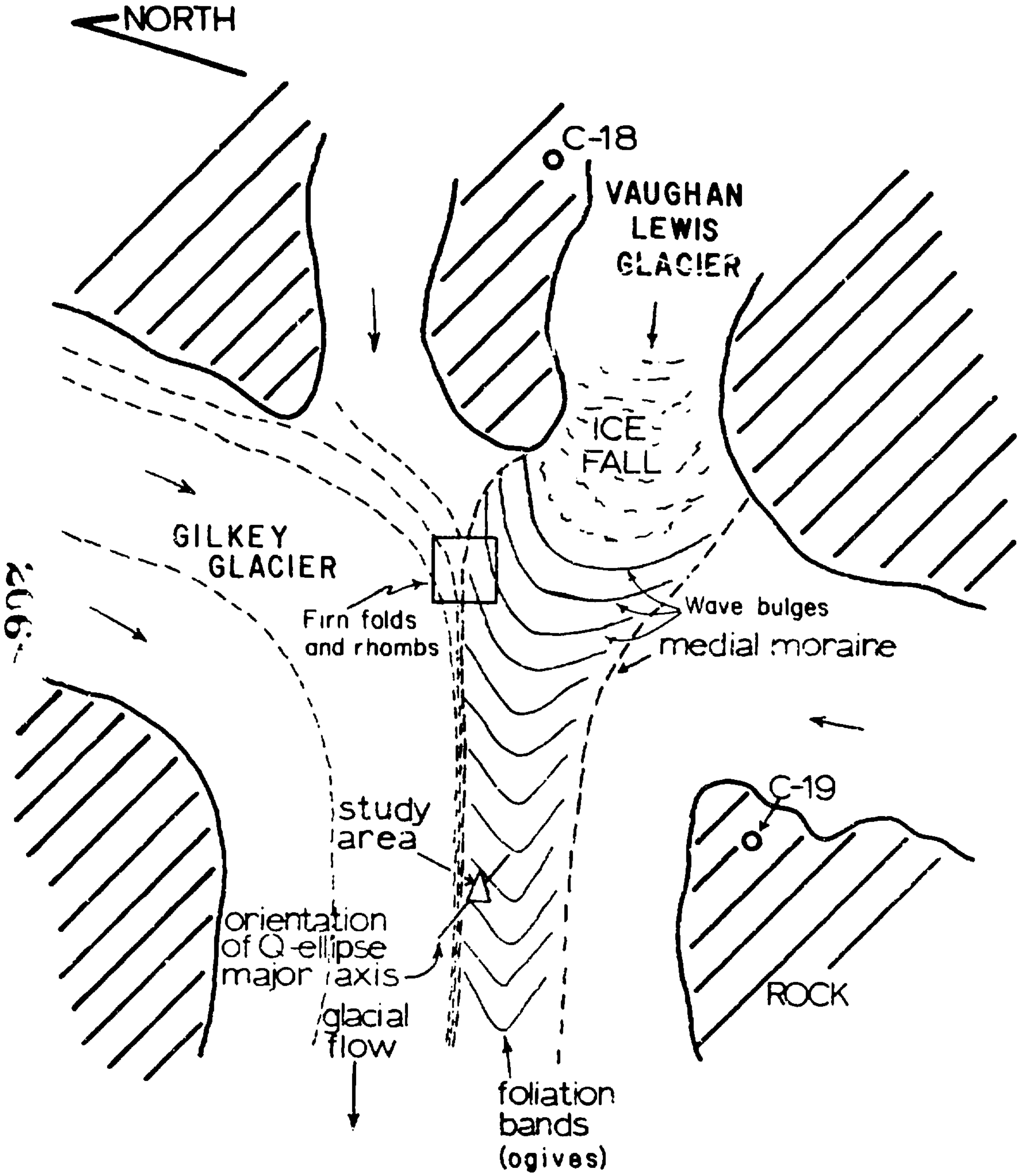


FIG 70a - Sketch map of Vaughan Lewis-Gilkey Glacier Research Area.

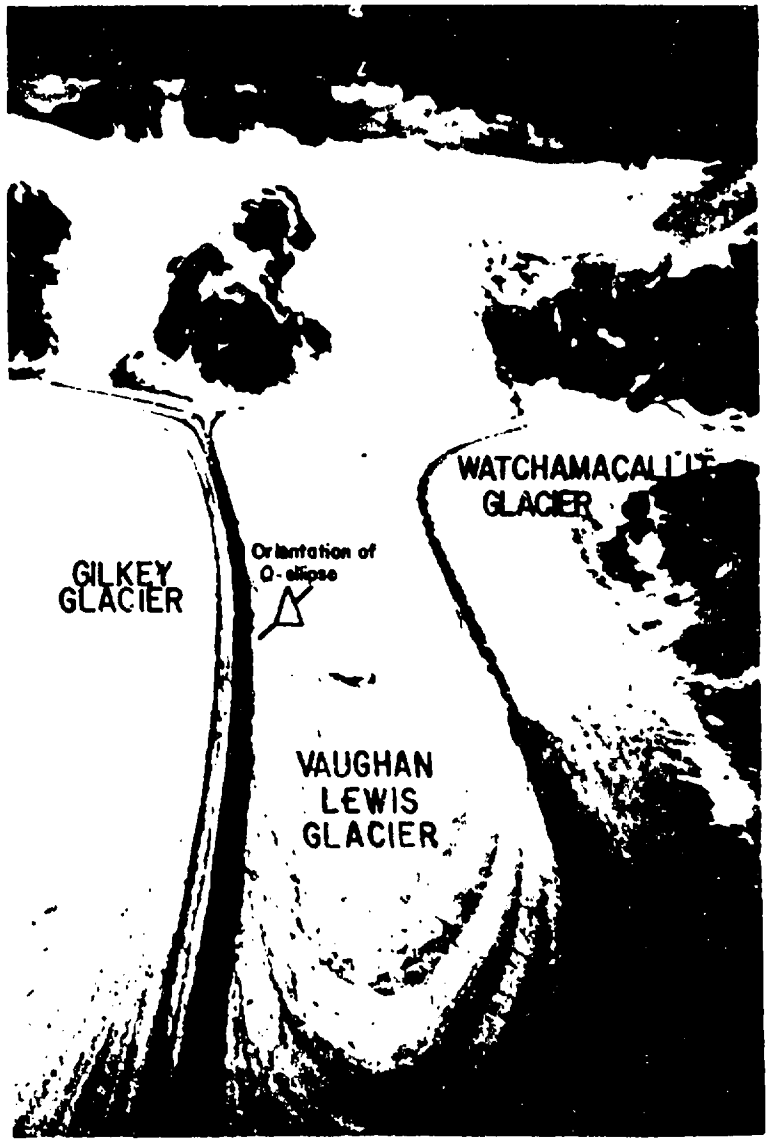


FIG 70b - Oblique aeriel view of Vaughan Lewis Glacier looking east.

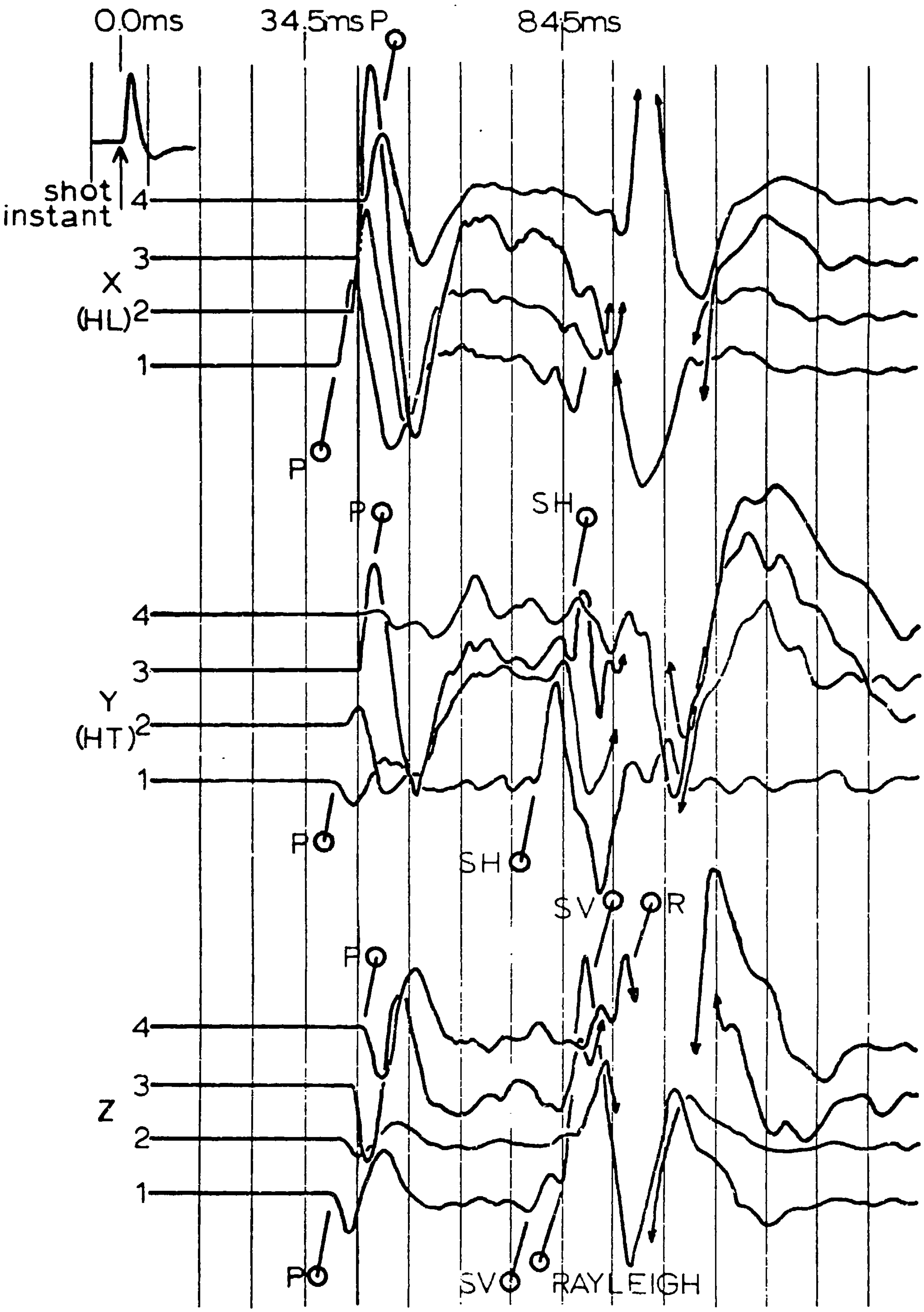


FIG. 71 GEOPHONE TRACES FROM SHOT 25, IN ANISOTROPISM STUDY, VAUGHAN LEWIS GLACIER

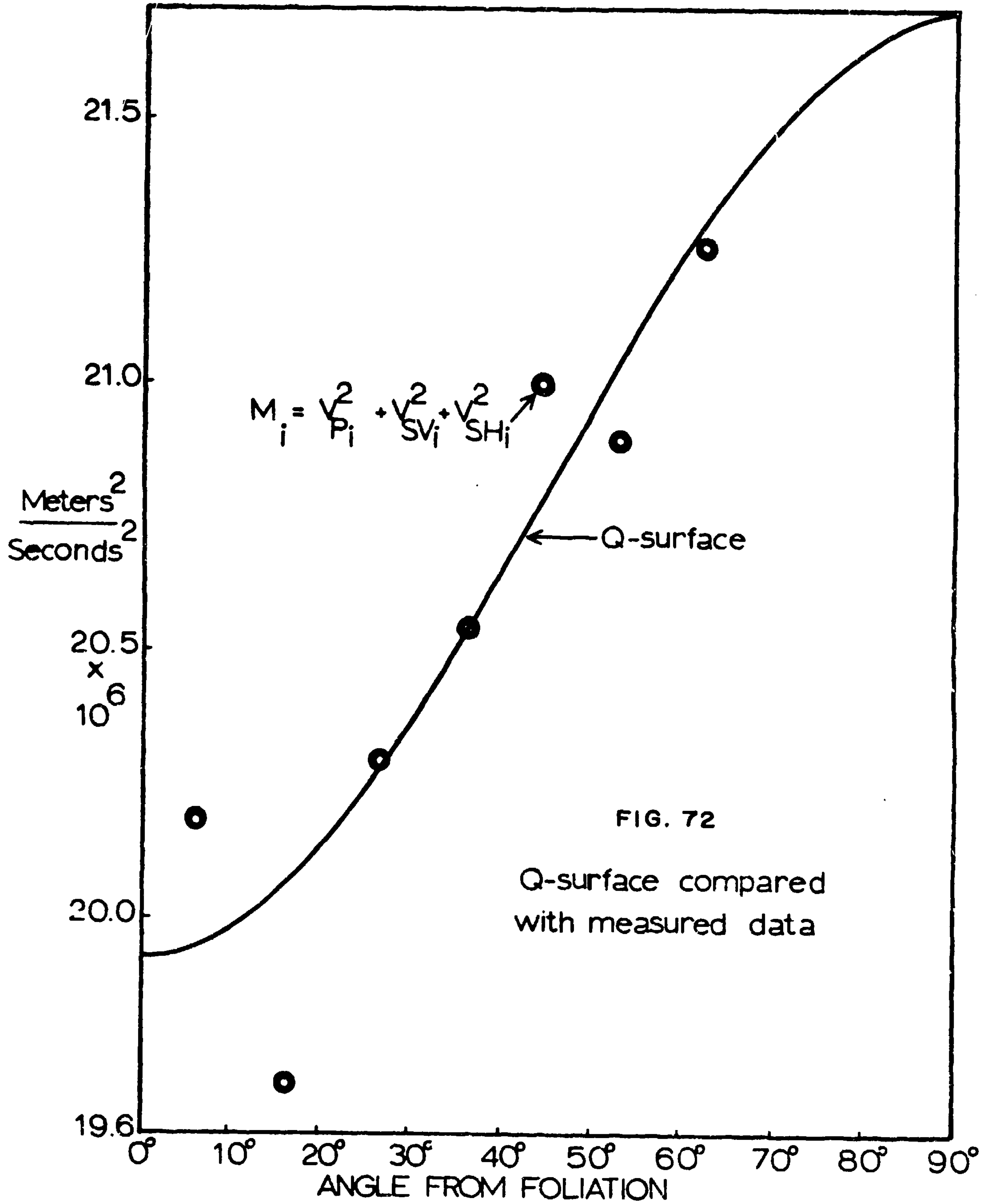
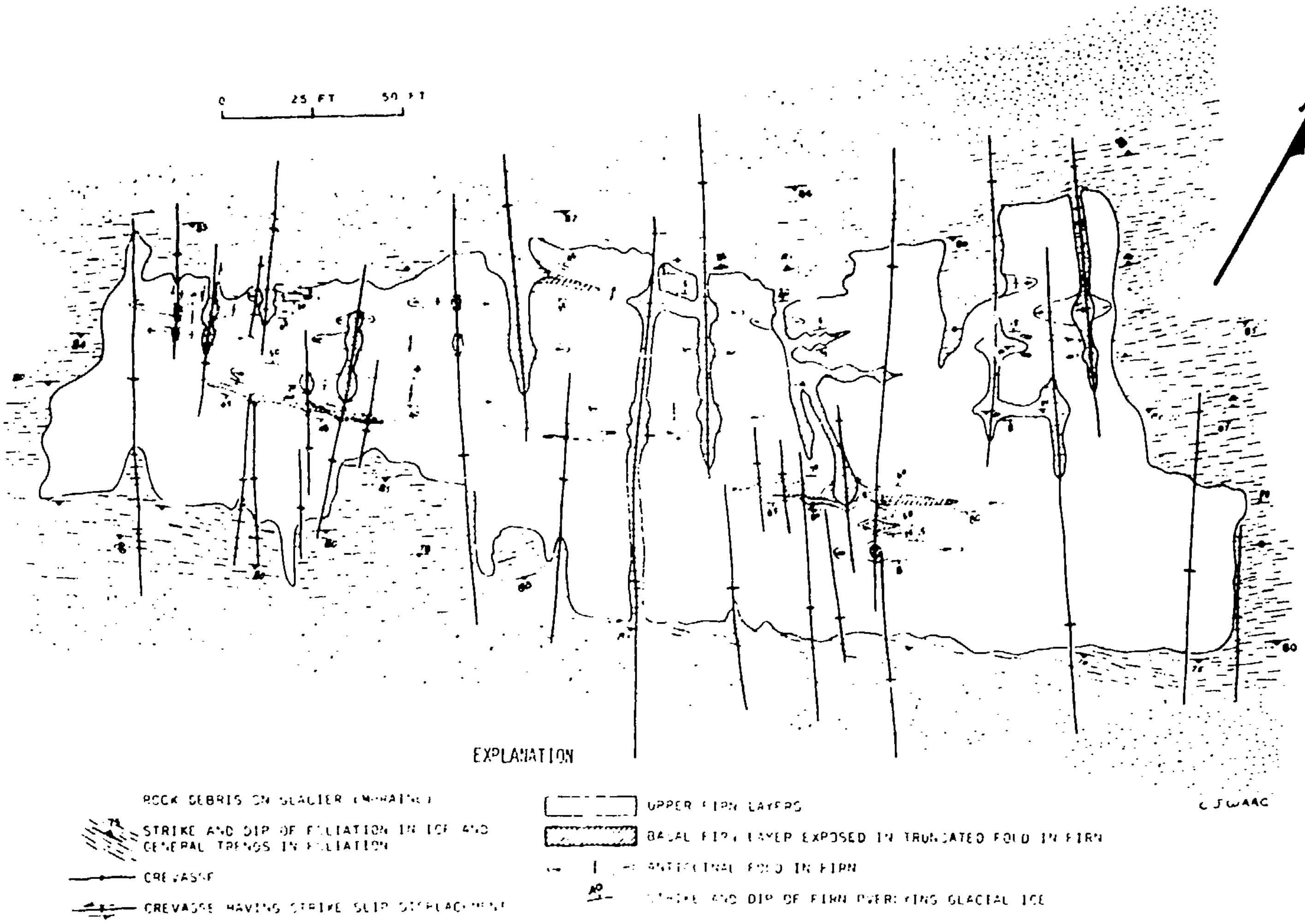
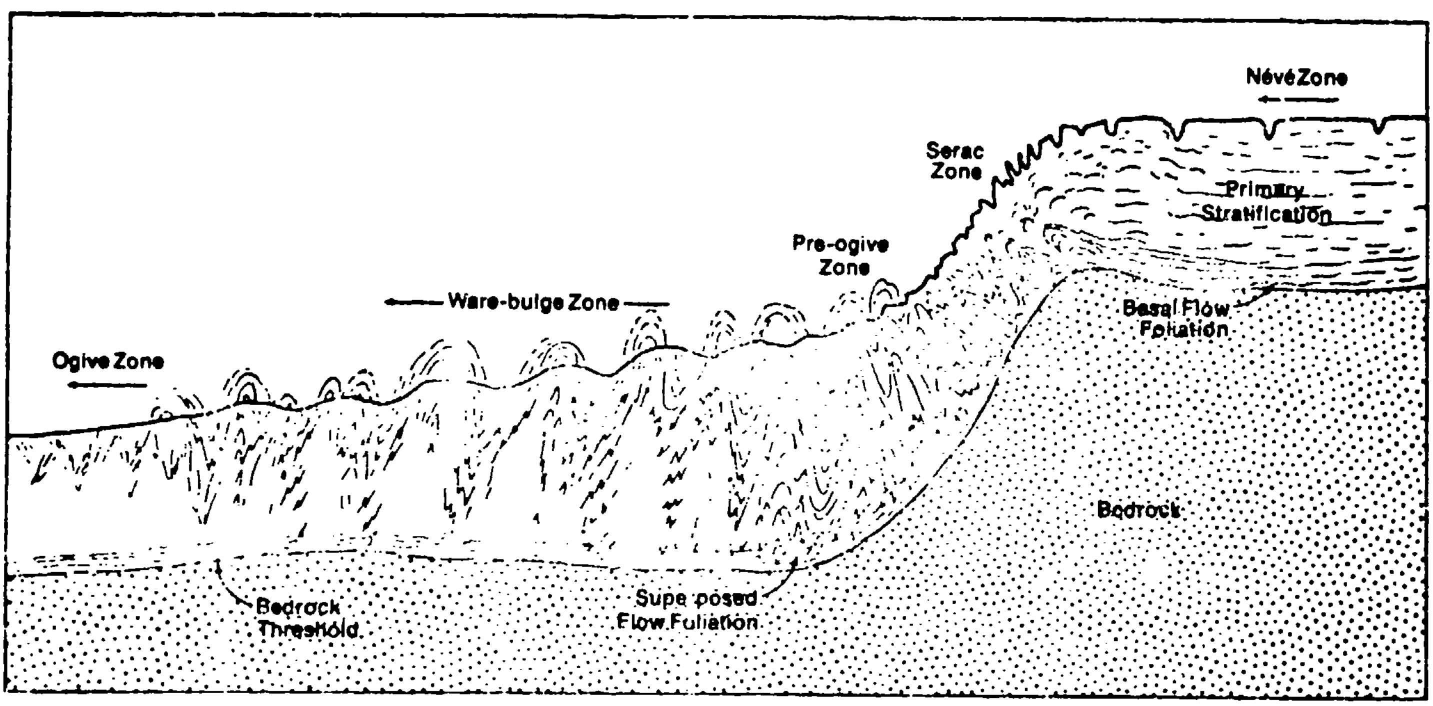


FIG. 72

Q-surface compared with measured data



(a) Plan view of structural detail in Gilkey Glacier confuence zone (after waag, 1975)



(b) Schematic of rotated foliation in and below icefall of Vaughan Lewis Glacier showing evolution of ogives (after Miller, 1968)

Fig. 73 Pertinent englacial structures of the Vaughan Lewis and Gilkey Glaciers.

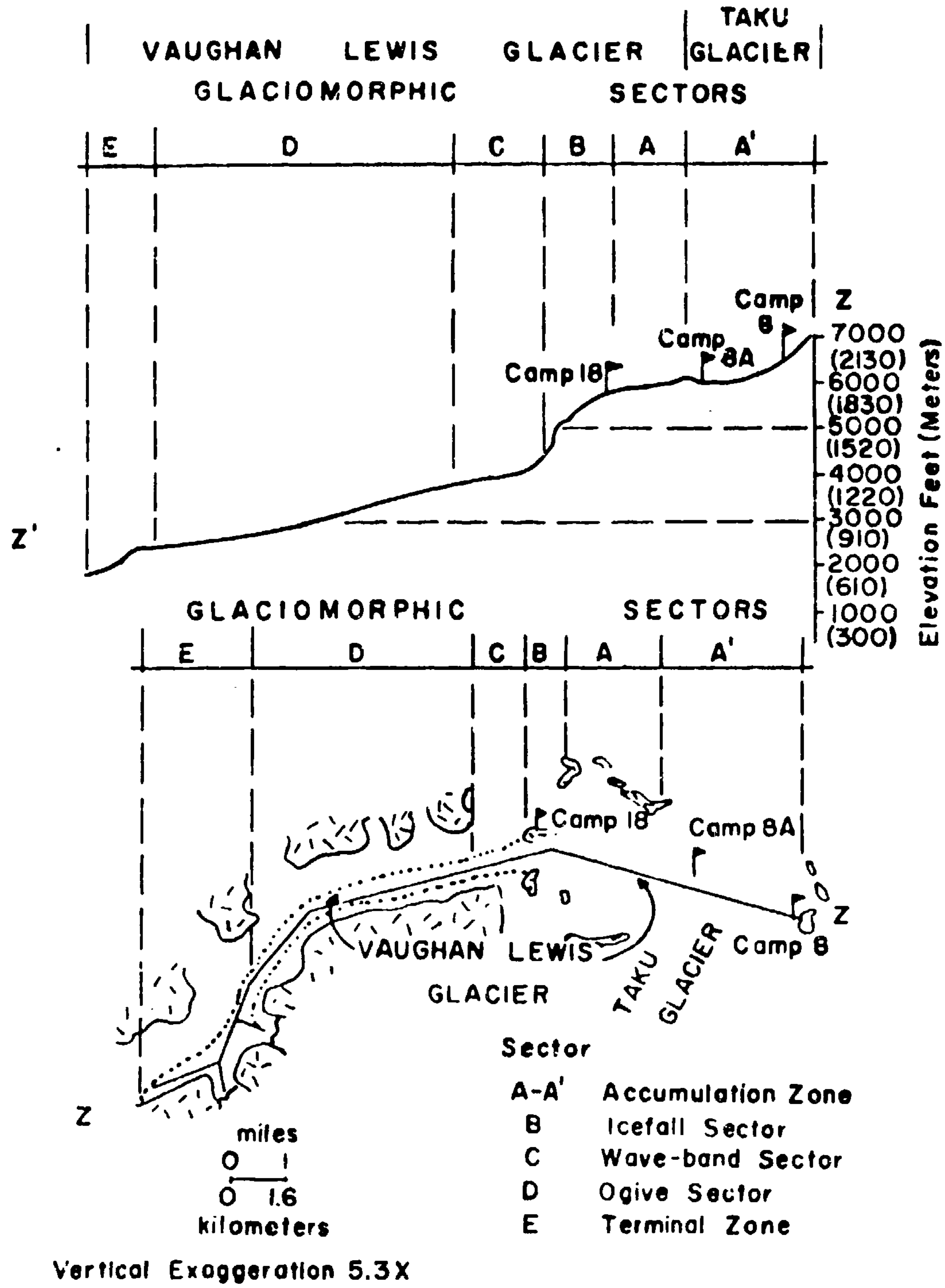


Fig. 74 Glaciomorphic sectors or regime zones of the Vaughan Lewis Glacier.

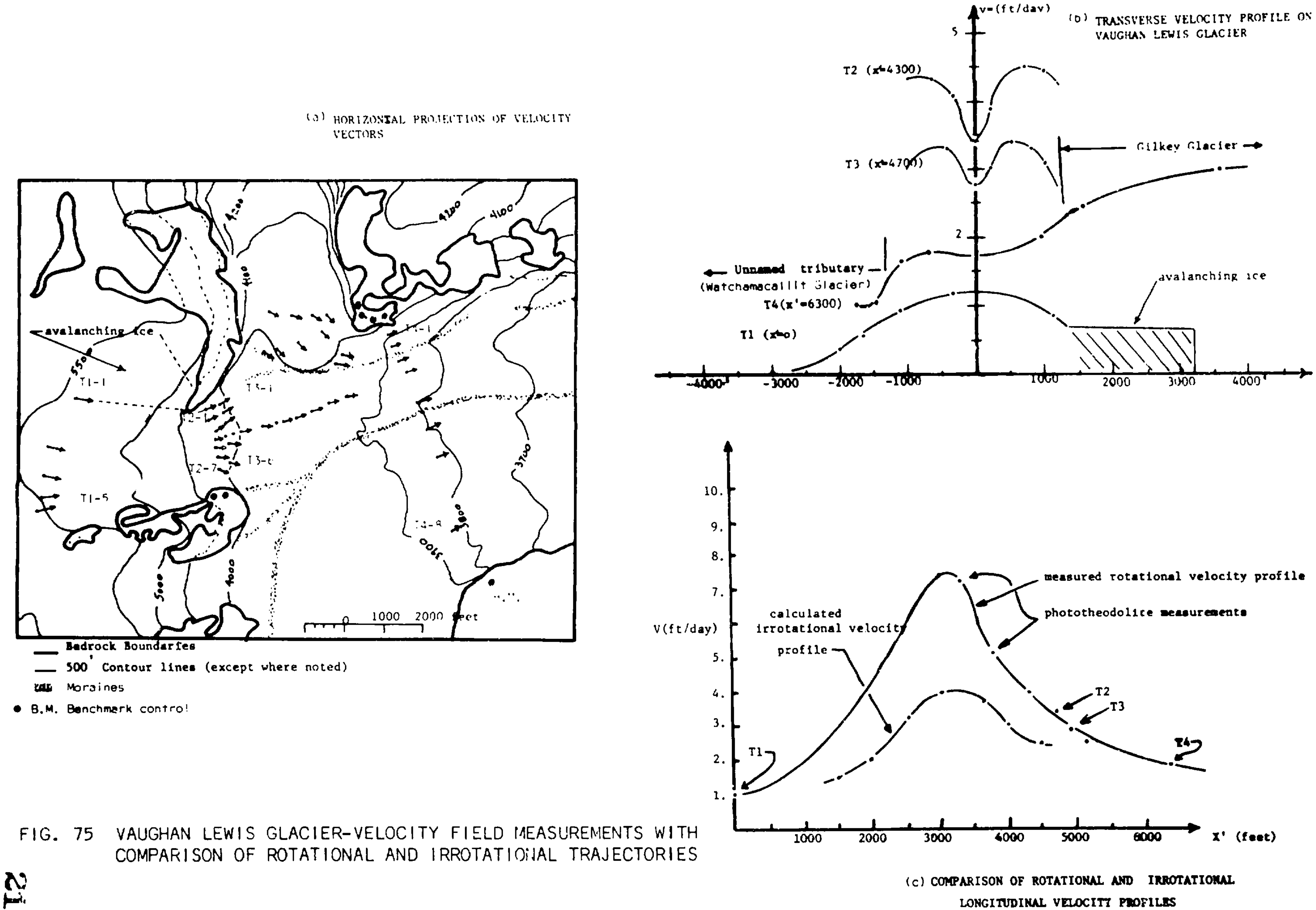


FIG. 75 VAUGHAN LEWIS GLACIER-VELOCITY FIELD MEASUREMENTS WITH COMPARISON OF ROTATIONAL AND IRROTATIONAL TRAJECTORIES

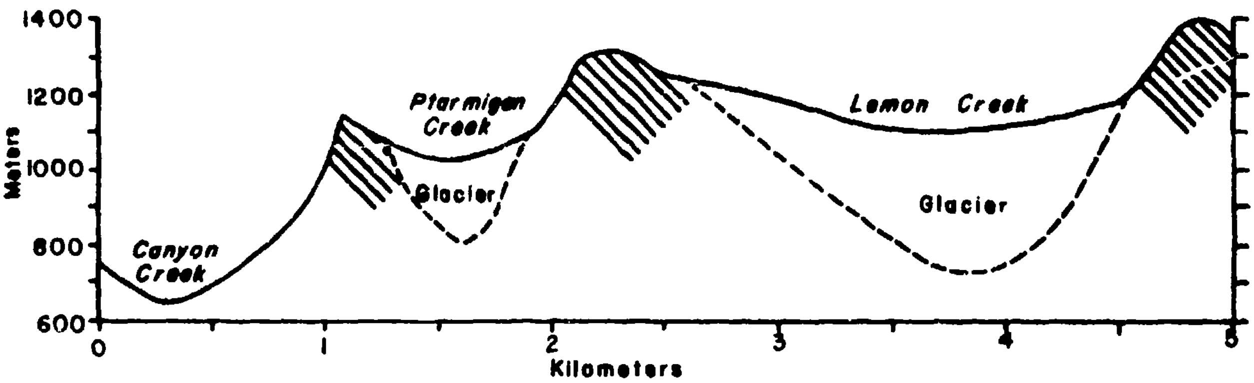


Fig. 76 - Diagrammatic representation of Lemon, Ptarmigan, and Canyon Creek valleys showing elements of bedrock control.

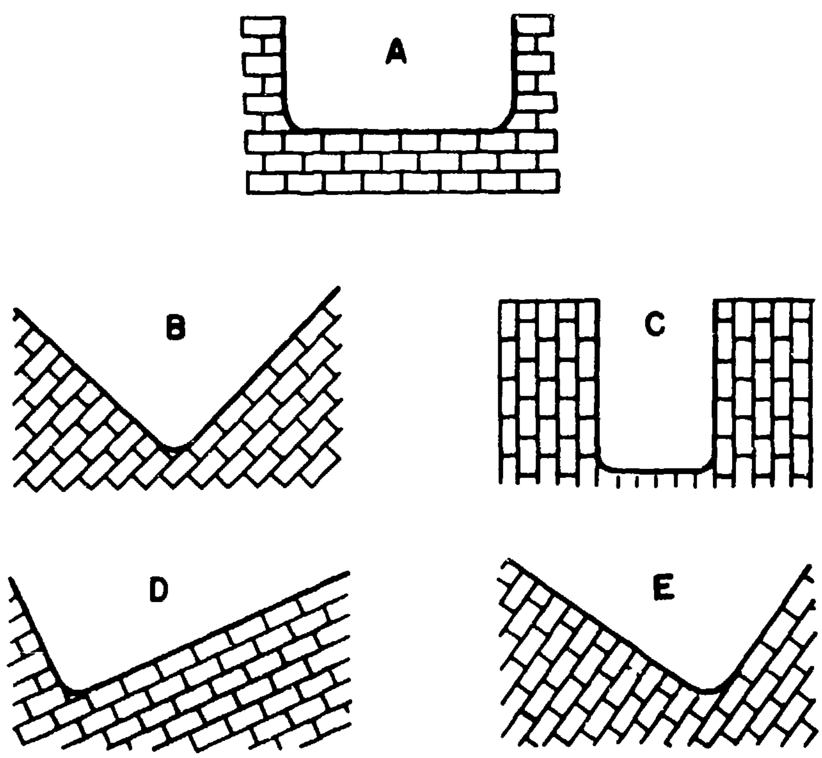


Fig. 77 - Model of Joint Geometries and Postulated Effects on Valley Shape (after Seppala).

(a) Talsekwe ice-dammed lake in July, 1951 just before flooding. View southwest to Devils Paw (8584')

(b) View east across Talsekwe Lake on July 14, 1975, the day before its Jukulhlaup, or sudden discharge beneath Talsekwe Glacier and down the Taku Valley.

Fig. 78 Two views of rugged terrain in southeastern sector of the Juneau Icefield on the Alaska - B.C. border (Photos M. M. Miller)

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* Represent papers in professional journals, in Symposia Proceedings, in press, or as Open File Research-in-Progress reports resulting from research dupported by this ARO-D grant.

** Represent theses also serving as reports on this research

APPENDIX A

I. Publications Resulting from this Contract Study

Publications and reports resulting from research supported by this Army Research Office contract are noted by asterisks in the Bibliography. These represent 18 papers in professional journals; 24 in Symposia Proceedings; 7 in press in various media; and 14 as Open File Research-in-Progress reports. Theses serving as reports are noted by double asterisks.

II. List of Theses and Advanced Degrees Earned by Participating Personnel

A. By participants supported in some way by the ARO-D grant funds.

Barry W. Prather (1972) M.S. degree, Dept. of Geology, Michigan State University
Thesis: Studies in Glacier Anisotropy on the Vaughan Lewis Glacier, Alaska.

Chester R. Zenone (1972) M.S. degree, Dept. of Geology, Michigan State University
Thesis: Glacio-hydrological Parameters of the Mass Balance of the Lemon Glacier, Alaska

Gordon Warner (1973) Ph.D. degree, Dept. of Metallurgy, Mechanics and Materials Science, Michigan State University
Thesis: The Measurement of Glacier Strain using Embedded Resistance Strain Gages (with special reference to investigations of the Ptarmigan and Cathedral Glaciers, Juneau Icefield, Alaska-B.C.)

Vernon K. Jones (1975) M.S. degree, Dept. of Geology, Michigan State University
Thesis: Contributions to the Geomorphology and Neogladal Chronology of the Cathedral Glacier System, Atlin Wilderness Park, British Columbia

Ann M. Tallman (1975) Ph.D. degree, Dept. of Geology, Michigan State University
Thesis: Glacial and Periglacial Geomorphology of the Fourth of July Creek Valley, Atlin Region, Cassiar District, N.W. British Columbia

B. By participants not supported by the ARO-D grant funds.

Susan Tomlinson (1973) M.S. degree, College of Forestry, Syracuse University
Thesis: A Dendrochronologic Analysis of the Arctic Shrub, Cassiope Mertensiana, on the Juneau Icefield, Alaska

Christopher P. Egan (1971) Ph.D. thesis, Dept. of Geology, Michigan State University.
Thesis: Contribution to the Late Neoglacial History of the Lynn Canal and Taku Valley Sector of the Alaskan Boundary Range.

C. Theses in Progress by Participating Personnel

Vernon K. Jones..Ph.D. in progress, Dept. of Agricultural Engineering, Michigan State University. Tentative thesis title:
Glacio-Climatic Trends and Global Climatic Change...A Concern for Agricultural Planning

Richard Warren..Ph.D. in progress, Dept. of Mechanical Engineering, Michigan State University. Tentative thesis title:
(On Glacier Flow Dynamics)..An N-Dimensional Matrix...Its Application to Anisotropic Materials

APPENDIX B

LIST OF SCIENTIFIC PERSONNEL INVOLVED IN THIS RESEARCH PROGRAM, 1971-74

Dr. Maynard M. Miller, Dept. of Geology, Michigan State University
Dr. James H. Anderson, Institute of Arctic Biology, University of Alaska
Stephen C. Andrews, Department of Geography, University of North Dakota
Robert A. Asher, Foundation for Glacier and Environmental Research, Seattle, Wa.
Dr. Hugh F. Bennett, Dept. of Geology, Michigan State University
Dr. Robert F. Black, Dept. of Geology, University of Connecticut
Lawrence Brush, Dept. of Geology, Harvard University
Dr. James E. Bugh, Dept. of Geology, State University of New York, Cortland
Dr. Gary Cloud, Dept. of Metallurgy, Mechanics and Materials Science, Michigan State University
William A. Dittrich, Joint Institute for Laboratory Astrophysics, University of Colorado
Dr. Egon Dorrer, Dept. of Surveying Engineering, University of New Brunswick, Fredericton, N.B., Canada
William Gawthrop, National Center for Earthquake Research, U.S. Geological Survey
Austin E. Helmers, Institute of Northern Forestry, U.S. Forest Service, Fairbanks, Alaska
Richard Heffernan, Foundation for Glacier and Environmental Research, Seattle, Wa.
Dr. William Hinze, Dept. of Geology and Geophysics, Purdue University
Vernon K. Jones, Dept. of Geology, Michigan State University
Dr. Richard Kellogg, Dept. of Physics, Luther College, Iowa
Dr. Gottfried Konecny, Professor of Photogrammetry, Technical University of Hanover, Germany
Charles Kreitler, Dept. of Geology, University of Texas, Austin
Dr. Edward E. Little, Foundation for Glacier and Environmental Research, Seattle, Wa.
William M. Lokey, U.S. Antarctic Research Program (NSF-USARP)
Charles Merry, Dept. of Surveying Engineering, University of New Brunswick, Fredericton, Canada
F. Nishio, Institute of Low Temperature Science, University of Hokkaido, Japan
Dr. Alfred C. Pinchak, Dept. of Fluid Mechanics, Case Western Reserve University
Barry W. Prather, Dept. of Geology, Michigan State University
Marianne See, Dept. of Botany, Smithsonian Institution
Dr. Matti Seppala, Dept. of Geography, University of Turku, Finland
Richard M. Shaw, Headquarters Exploration, Exxon, Denver, Colorado
Dr. Aylmer H. Thompson, Dept. of Meteorology, Texas A and M. University
Dr. Ann M. Tallman, Dept. of Geology, Smith College
Dr. Gorow Wakahama, Institute of Low Temperature Science, University of Hokkaido, Japan
William Walkotten, Institute of Northern Forestry, U.S. Forest Service, Juneau, Alaska
Richard Warren, Dept. of Geology, Michigan State University
Dr. Gordon G. Warner, General Motors Institute, Flint, Mi.
Dr. Charles Waag, Dept. of Geology, Georgia State University
Chester Zenone, U.S. Geological Survey, Anchorage, Alaska

Additional Research Affiliates, not supported by ARO-D

Dr. Heinz Slupetsky, Dept. of Geography, University of Salzburg, Austria
Dr. Douglas N. Swanston, U.S. Forest Service, Forestry Sciences Lab., Corvallis, Or.
Hans-Dirk Jannsen, Universitat Karlsruhe, Germany (Surveying and Photogrammetry Dept.)
Karsten Jacobsen, Technische Universität, Hannover, Germany (Surveying Engineering Dept.)
Jon Kostoris, Dept. of Geology, Michigan State University
Susan Tomlinson, College of Forestry, State University of New York, Syracuse

APPENDIX C

RELATIVE MEAN SUNSPOT NUMBERS (Rz)

As supplied by Swiss Federal Observatory, Zurich, Switzerland, Published in Monthly Weather Review, Vol. 30 et seq.; and also from World Data Center for Solar-Terrestrial Physics, NOAA*

Year	Mean (Rz)	Year	Mean (Rz)	Year	Mean (Rz)
1749	80.9	1791	66.6	1833	8.5
1750	83.4	1792	60.0	1834	13.2
1751	47.7	1793	46.9	1835	56.9
1752	47.8	1794	41.0	1836	121.5
1753	30.7	1795	21.3	1837	138.3
1754	12.2	1796	16.0	1838	103.2
1755	9.6	1797	6.4	1839	85.8
1756	10.2	1798	4.1	1840	63.2
1757	32.4	1799	6.8	1841	36.8
1758	47.6	1800	14.5	1842	24.2
1759	54.0	1801	34.0	1843	10.7
1760	62.9	1802	45.0	1844	15.0
1761	85.9	1803	43.1	1845	40.1
1762	61.2	1804	47.5	1846	61.5
1763	45.1	1805	42.2	1847	98.5
1764	36.4	1806	28.1	1848	124.3
1765	20.9	1807	10.1	1849	95.9
1766	11.4	1808	8.1	1850	66.5
1767	37.8	1809	2.5	1851	64.5
1768	69.8	1810	0.0	1852	54.2
1769	106.1	1811	1.4	1853	39.0
1770	100.8	1812	5.0	1854	20.6
1771	81.6	1813	12.2	1855	6.7
1772	66.5	1814	13.9	1856	4.3
1773	34.8	1815	35.4	1857	22.8
1774	30.6	1816	45.8	1858	54.8
1775	7.0	1817	41.1	1859	93.8
1776	19.8	1818	30.4	1360	95.7
1777	92.5	1819	23.9	1861	77.2
1778	154.4	1820	15.7	1862	59.1
1779	125.9	1821	6.6	1863	44.0
1780	84.8	1822	4.0	1864	47.0
1781	68.1	1823	1.8	1865	30.5
1782	38.5	1824	8.5	1866	16.3
1783	22.8	1825	16.6	1867	7.3
1784	10.2	1826	36.3	1868	37.3
1785	24.1	1827	49.7	1869	73.9
1786	82.9	1828	62.5	1870	139.1
1787	132.0	1829	67.0	1871	111.2
1788	130.9	1830	71.0	1872	101.7
1789	118.1	1831	47.8	1873	66.3
1790	89.9	1832	27.5	1874	44.7

* World Data Center for Solar-Terrestrial Physics
Environmental Data Service, NOAA
Boulder, Colorado, USA 80302 (313) 499-1000 Ext. 6467

Year	Mean (Rz)
1875	17.1
1876	11.3
1877	12.3
1878	3.4
1879	6.0
1880	32.3
1881	54.3
1882	59.7
1883	63.7
1884	63.5
1885	52.2
1886	25.4
1887	13.1
1888	6.8
1889	6.3
1890	7.1
1891	35.6
1892	73.0
1893	84.9
1894	78.0
1895	64.0
1896	41.8
1897	26.2
1898	26.7
1899	12.1
1900	9.5
1901	2.7
1902	5.0
1903	24.4
1904	42.0
1905	63.5
1906	53.8
1907	62.0
1908	48.5
1909	43.9
1910	18.6
1911	5.7
1912	3.6
1913	1.4
1914	9.6
1915	47.4

Year	Mean (Rz)
1916	55.4
1917	103.9
1918	80.6
1919	63.6
1920	38.7
1921	24.7
1922	14.7
1923	5.8
1924	16.7
1925	44.3
1926	63.9
1927	69.0
1928	77.8
1929	65.0
1930	35.7
1931	21.2
1932	11.1
1933	5.7
1934	8.7
1935	36.1
1936	79.7
1937	114.4
1938	109.6
1939	88.8
1940	67.8
1941	47.5
1942	30.6
1943	16.3
1944	9.6
1945	33.1
1946	92.5
1947	151.5
1948	136.2
1949	135.1
1950	83.7
1951	69.4
1952	31.4
1953	13.8
1954	4.4
1955	37.9
1956	141.7

Year	Mean (Rz)
1957	189.8
1958	184.6
1959	158.7
1960	112.3
1961	53.9
1962	37.6
1963	27.9
1964	10.2
1965	15.1
1966	46.9
1967	93.7
1968	105.9
1969	105.6
1970	104.2
1971	66.6
1972	68.9
1973	38.0
1974	34.5
1975	18.3(proj.)

APPENDIX D

STREAM GAGING METHODS USED IN THE JUNEAU ICEFIELD RESEARCH

During the course of the glacio-hydrological investigations carried out on the Juneau Icefield between 1951 and 1975, including the three years of this ARO grant, standard stream gaging methods have been employed. These are based on U.S. Geological Survey procedures outlined in the following special reports: U.S. Geological Survey (1968) Water resources investigations of the U.S. Geological Survey. Book III. Applications of Hydraulics. Ch. A-6, General procedures for gaging streams, by Carter, R.W. and Davidian, J. Also, Ch. A-7, Stage measurements at gaging stations, by Buchanan, T.J. and Somers, W.P.

Briefly, the methods and equipment used in the ARO studies on Lemon Creek, Ptarmigan Creek (Camp I, Trail Cabin, via Camps 17 and 17A) and on Cathedral (Elixir) Creek near Camp 29 were as follows.

Stream discharge is the volume rate of water flow and includes sediments in suspension or dissolved in it. The discharge is given in ft^3/sec. The current meter method of measuring discharge was used at all three sites. The formula Q = (av) represents the products of the partial areas of the stream cross-section and its average velocities. (Note the following graph denoting the midsection method of computing cross-sectional areas.) In this:

(Q) is total discharge
(a) is a partial cross-sectional area
(v) is the corresponding mean velocity

The area, considered to be made up of a sum of rectangular sections, extends from half the horizontal distance from the previous observation point to half the distance to the next, and from the surface of the stream to the depth sounded. In order to obtain the mean vertical distribution of velocity, velocities are sampled at each observation point with a standard current meter. In this study a Type AA vertical axis Price Meter was used. (Where small stream flow is measured, as in Cathedral Creek at low water times, a Pygmy type was employed.) The following equation is used to calculate the partial discharge:

$$qx = V_x \left[\frac{(b_x - b(x-1)}{2} + \frac{(b(x+1) - bx)}{2} \right] dx$$
$$= V_x \left[\frac{b(x+1) - b(x-1)}{2} \right] dx$$

In this,

q_x = discharge through partial section x
V_x = mean velocity at location x
bx = distance from initial point to location x
b(x-1) = distance from initial point to preceding location
b(x+1) = distance from initial point to next location
dx = depth of water at location x

The current meter, by earphone connection, allows one to count the number of revolutions of the rotor during a given period of time. From the number of revolutions and time, the velocity of flow is found. Frequency is measured by electrical impulses produced in a contact chamber and giving audible clicks in the headphone. The time intervals are recorded by stopwatch.

In the Ptarmigan and Cathedral Creek gaging, a top-setting wading rod was used for sounding. The rod was carefully placed in the stream so that the base plate was stable on the stream bed. The depth was read on the graduated main rod and the velocity measured with a Price meter attached to the rod.

Making the discharge measurements at a cross-section requires a mean velocity in each selected vertical, but the current meter determines velocity at a point. The method used on the Lemon Creek was the two-point method. Observations were made in each vertical sounding at points 0.2 and 0.8 of the total distance to the bed from the surface. By averaging this, the mean velocity was determined. This method was used with depths of 2.5 feet or more. When the depth was between 0.3 and 2.5 feet (as in Ptarmigan and Cathedral Creeks), a second method was invoked which made observations of velocity at 0.6 of the total depth below the surface. This was considered to be the mean velocity.

The stage of a stream is defined as the height of the water surface above an established datum plane. Measurements of this level are essential to determinations of stream discharge. The method used to sense the water level or stage was a float in a stilling well, inside of a weather protection shelter. Intake pipes connected the stream to the gage. The Lemon Creek site had a stilling well placed directly in the stream. In the Ptarmigan and Cathedral Creek surveys, the well was placed under the bank of the stream. All three stilling wells were covered to provide adequate protection to the float and prevent the direct influence of winds on variations in water level.

The instruments used were Stevens A water-stage recorders, with continuous strip-charts providing a quite satisfactory stage record. For the scale on the graphic record, a 1:6 gage height scale and a time scale of one day equalling 2.4 inches were chosen. To insure accuracy, comparative readings were taken on the inside and the outside gages at each site visit. This also helped to detect any malfunctions of the recording system.

A discharge rating curve defines discharge as a function of stage, slope, rate of change of stage, or other variables. The Lemon Creek discharge ratings consisted of simple relations between stage and discharge. Several discharge measurements are essential to define the changing relationship. If the control changes, a new discharge rating is needed. All three creeks surveyed are vulnerable to changes in stream channel due to the influence of ice, deposition, erosion and general changes in bed roughness, thus requiring repeated gaging several times each season.

The data are plotted on logarithmic paper. From a graphical examination of the curve, plotted as a straight line, the stage-discharge relations of each creek were found. Included in this appendix, as an example, is the Lemon Creek Discharge rating curve(after Guigné, 1974).*

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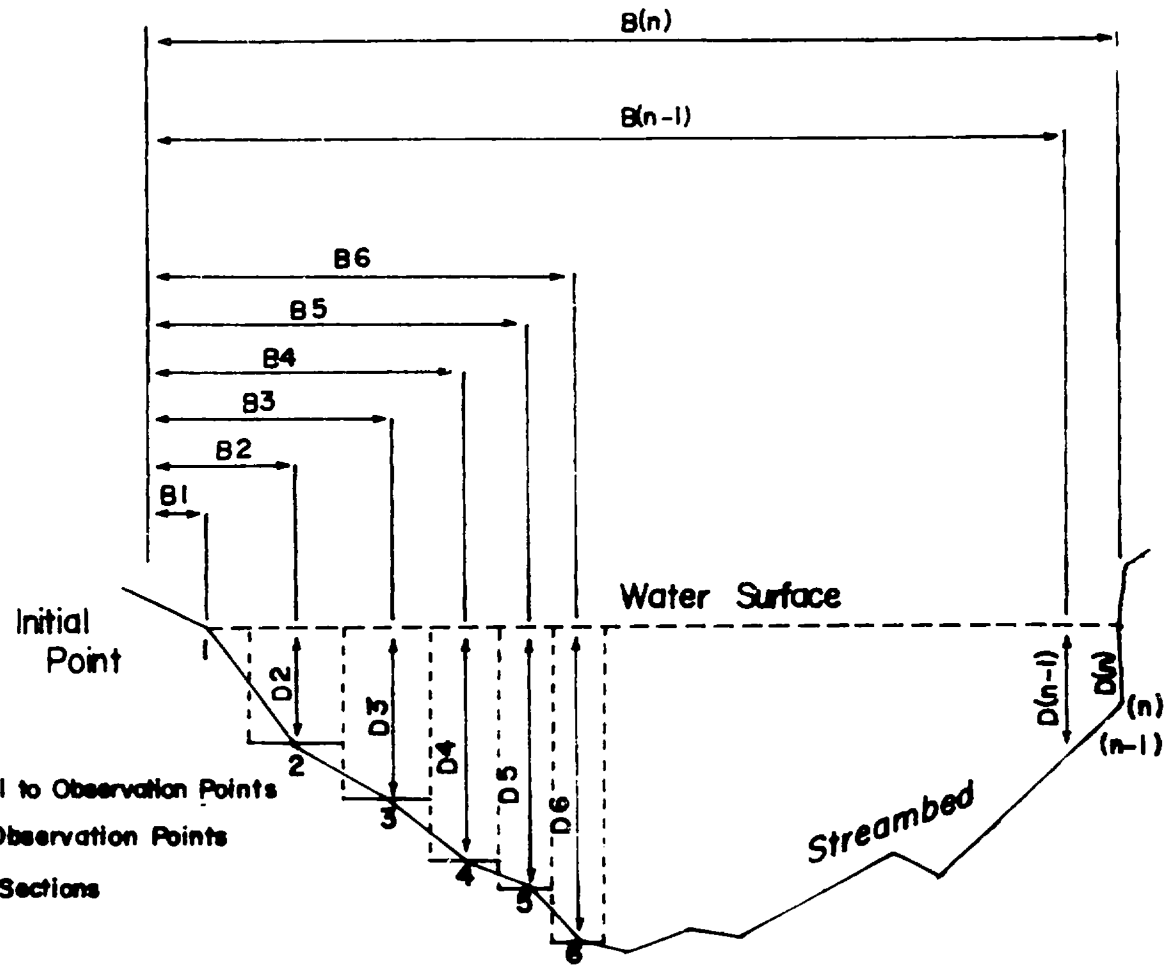
* Guigné, Jacques Yves (1974) Hydrological Studies on the Lemon, Ptarmigan and Cathedral Glacier Systems, 1974. Open File Report, Juneau Icefield Research Program, 1974 season.

MIDSECTION METHOD of COMPUTING CROSS-SECTION AREA for DISCHARGE MEASUREMENTS (definition sketch)

230<



EXPLANATION

1,2,3...n Observation Points

B1, B2..B(n) Distance from Initial to Observation Points

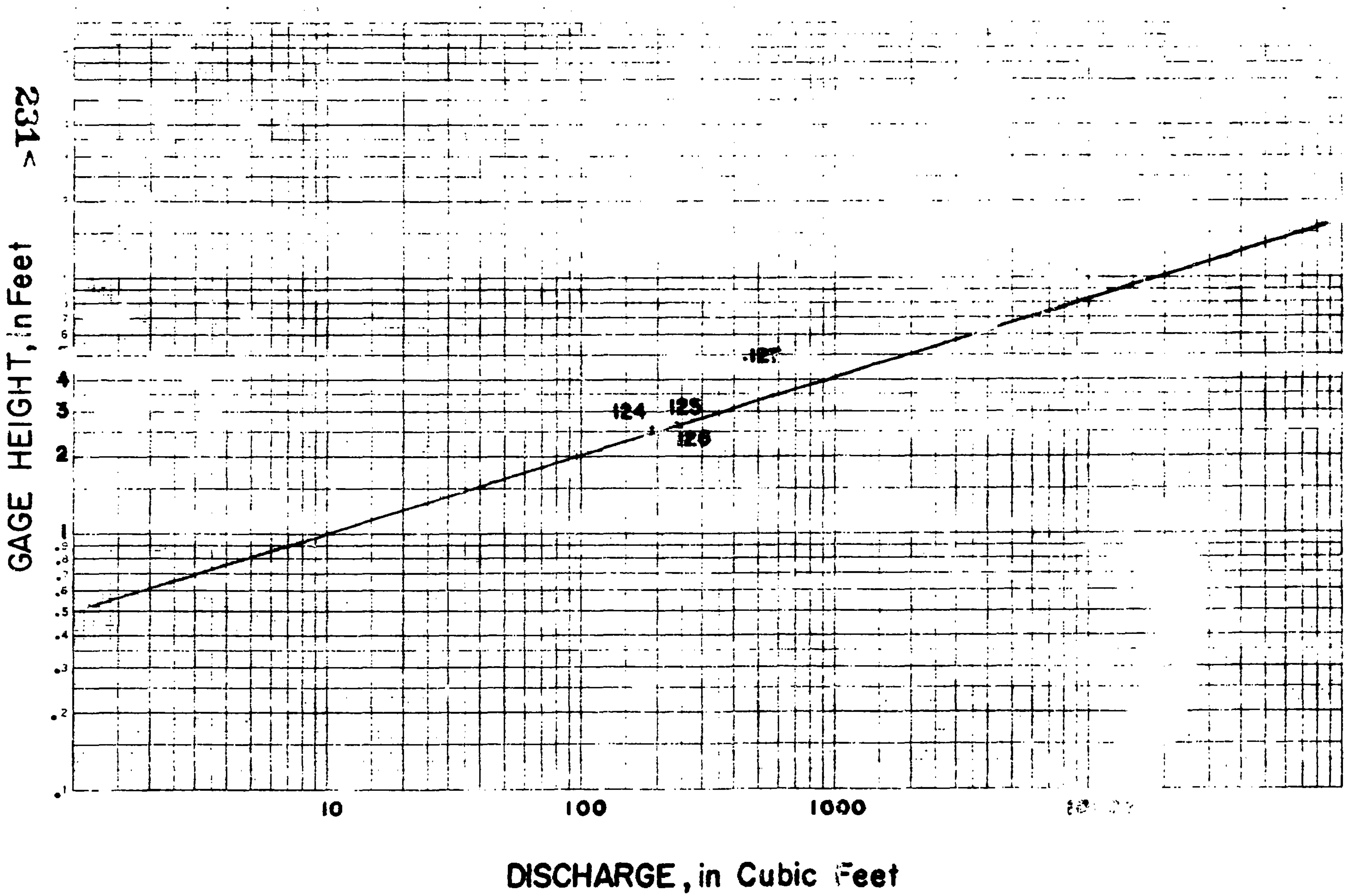
D1, D2..D(n) Depth of Water at Observation Points

Dash-Line..Boundary of partial Sections

(based on DISCHARGE MEASUREMENTS at GAGING STATIONS by Buchanan and Somers U.S.G.S.)

J.Y. GUIGNE 2/75

DISCHARGE RATING CURVE for LEMON CREEK (7/'74 and 8/'74)



NOTE: Hydraulic Measurements and Stage-Discharge Relation Computed by J.Y. GUIGNÉ (2/'75)

E-1

APPENDIX E

JUNEAU ICEFIELD RESEARCH PROGRAM — GLACIO-METEOROLOGY RECORD

MONTHLY SUMMARY WEATHER FORM

Station or Camp No. **17, Lemon Gl.** Date **June 16-July 23 1973** Elevation **4200 ft.**

Date	Max. Temp.	Min. Temp.	Mean Temp.	Av. Rel. Humidity	Av. Cloud Cover	Total Precip. -Rain	Total Snow	Av. Wind Speed 21 H	Av. Wind Speed Day	Av. Wind Speed Night	Av. Wind Direction	Days Fog	Gross Surf. Ablation	Duration Sunshine	Radiation
1	49	38	43.5		8	T			3		N			5.75	
1	50.5	41	45.8		9	T			3		S			1.5	
18	40	34	37	83	9	0.11			5		SE			2.5	
19	40.5	33	37.3	91	8	0.09	T		6		SE			1.75	
20	41	32	36.5	88	4	0.01			1		E		88.	14.75	
21	42	36	39	99	10	0.22			17		SE		96	0.25	
22	38	29	3[illegible].3	90	9	T			8		E		98	4.5	
	42	37	39.5	91	10	0.04			15		ESE		100	0	
	40	29	34.5	100	10	0.59			15		SE		107	0	
	37.5	33	35.3	100	10	0.55	T		3		SE		111	0	
	37.5	31	34	89	9	0.23	T		5		SE		112	2.25	
	39.4	31	35.2	99	10	0.15			6		SE		116	0.25	
	36	32	34.2	100	10	0.15	T		7		SW		119	0	
	38	33	35.5	100	10	0.22			5		W		122	0	
	38	32	34.8	100	10	0.26			2		WSW		124	0	
1	36.5	32	34.2	98	10	0.15	T		9		SE		127	0	
2	39	32	35.3	98	10	0.19	T		16		E		130	0	
3	38.5	34	36	98	9	0.28	T		17		SE		135	1.25	
4	42.5	35	38.8	96	9	0.14			11		SE		139	0	
5	36.6	31	33.9	100	10	0.20	0.9		9		W		137	0	
	46	32	39	93	8	0.01			2		E		143	0	
	43	34	38.5	96	10	0.03			3		S		146	0	
8	45	36	40.7	95	10	0.17			3		SW		150	0	
9	44.5	39	41.5	93	9	0.31			7		SW		156	0.1	
10	48.8	37	42.9	80	6	0.01			9		NE		159	10.5	
11	40.5	35	37.5	97	10	1.45	T		12		SE		99	0	
12	39	34	36.5	99	10	0.76			31		SE		105	0	
13	40	34	37	93	10	1.04	T		27		S		115	0	
14	37	32	34.5	94	10	0.14	T		11		W		116	0	
15	37	33	35	95	10	0.03			8		NW		119	0	
16	51	32	41.5	71	4				6		W		123	11.5	
17	46	39	42.5	95	9	0.10			8		SSE		127	0	
18	46.5	41	44	100	10	0.78			15		SE		136	0	
19	41	38	39.3	100	10	0.61			13		SE		142	0	
20	40	32	35.8	94	10	0.85	0.4		8		SW		142	0	
21	37	32	34.5	99	10	0.18			11		W		29	0	
22	40	33	36.3	94	10	0.03			5		NW		31	0	
23	50.8	34	42.5	49	8	0.11			5		N		36	0	
C	1576	1268	1432	3557	348	10.19								56.85	
M	41.46	33.3	37.7	93.25	9										

R... Minimum temperatures and gross surface ablation figures rounded to nearest whole number.

APPENDIX F

FREE WATER CONTENT MEASURING PROCEDURES ON THE JUNEAU ICEFIELD

A. Description of Measuring Principle

Free water content of snow and firn means the ratio or proportion of liquid water in wet snow or firn when the snow or ice temperature is below 0°C (32°F). Unless we are dealing with super-cooled zones in sea-ice or a glacier there is no liquid water present.

The mass of wet snow (Ms) is comparised of both water mass (Mw) and ice mass (Mi), so the ratio of free water content (FWC) in wet snow must be represented as a percentage. Thus:

$$FWC = \frac{Mw}{Ms} \times 100 \qquad (\%) \qquad (1)$$

The percentage of free water content in wet snow can be obtained by non-calorimetric as well as calorimetric methods. The non-calorimetric method used on the Juneau Icefield Research Program has been described by Miller (1954).

The calorimetric method as used on the 1972 Juneau Icefield Research Program was carried out as follows. The calorimeter is shown in Figure

The procedures are as follows: Put hot water at temperature T_1 (mass is W) in the container 109-A* and wet snow (Ms) in the container 109-B, which are covered with a thermal insulator. Then both containers are connected and the wet snow (firn) is mixed with hot water and melted into water at temperature T_2. The following equation pertains:

$$(W + W_A)\ T = L\ Mi + (Ms + Wb)\ T2 \qquad (2)$$

Here W_A and Wb are constants depending on the calorimeter and on the water equivalent heat capacity. The factor T is T1 = T2 and L is the latent heat of fusion or of ice crystallization (79.6 calories per gram).

If we measure the values of W, T1, T2 and Ms, we can get the mass of ice (Mi) in the wet snow (firn) from equation (2).

On the other hand, wet snow or firn (Ms) is:

$$Ms = Mi + Mw \qquad (3)$$

So equation (1) and (3) gives

$$FWC = \left(1 - \frac{Mi}{Ms} \right) \times 100 \qquad (\%) \qquad (4)$$

---

* Institute of Low Temperature Science, University of Hokkaido, calorimeter, as used on the Juneau Icefield Research Program, 1971-75

Measurement Procedures

(1) Prepare hot water, between 55° and 60° C, in a pot on the stove.

(2) Pour the hot water into a plastic pot.

(3) Fill the blue bottle with the hot water, between 53° and 58° C.

(4) Insert proper thermometer into the blue bottle, until the rubber is reached. In this process, if a little hot water is spilt out by inserting the thermometer, this can be ignored. Next, measure the hot water temperature, so you can get the mass of hot water from Table I (at end of this appendix). One must be careful not to squeeze the blue bottle because this will preclude getting the correct volume of hot water.

(5) Transfer the hot water into the container 109-A with care.

(6) Leave the container, 109-A, covered with a thermal insulator.

(7) Collect wet snow (or firn) of 100-120 grams and put into the container, 109-B. It is recommended that between 210 and 230 grams be collected, including the weight of 109-B. This should be done quickly and carefully. Wear thick gloves, for the snow or firn will be melted by warm hands. The container must be cooled to 0°C. (Put the container, 109-B, into wet snow in advance.)

(8) Measure the weight of wet snow (firn) with the dead-weight of the container (without stopper).

(9) Shake 109-A to agitate the hot water and then read the initial temperature of the hot water.

(10) Connect 109-A and 109-B and then mix the hot water with the wet snow (firn).

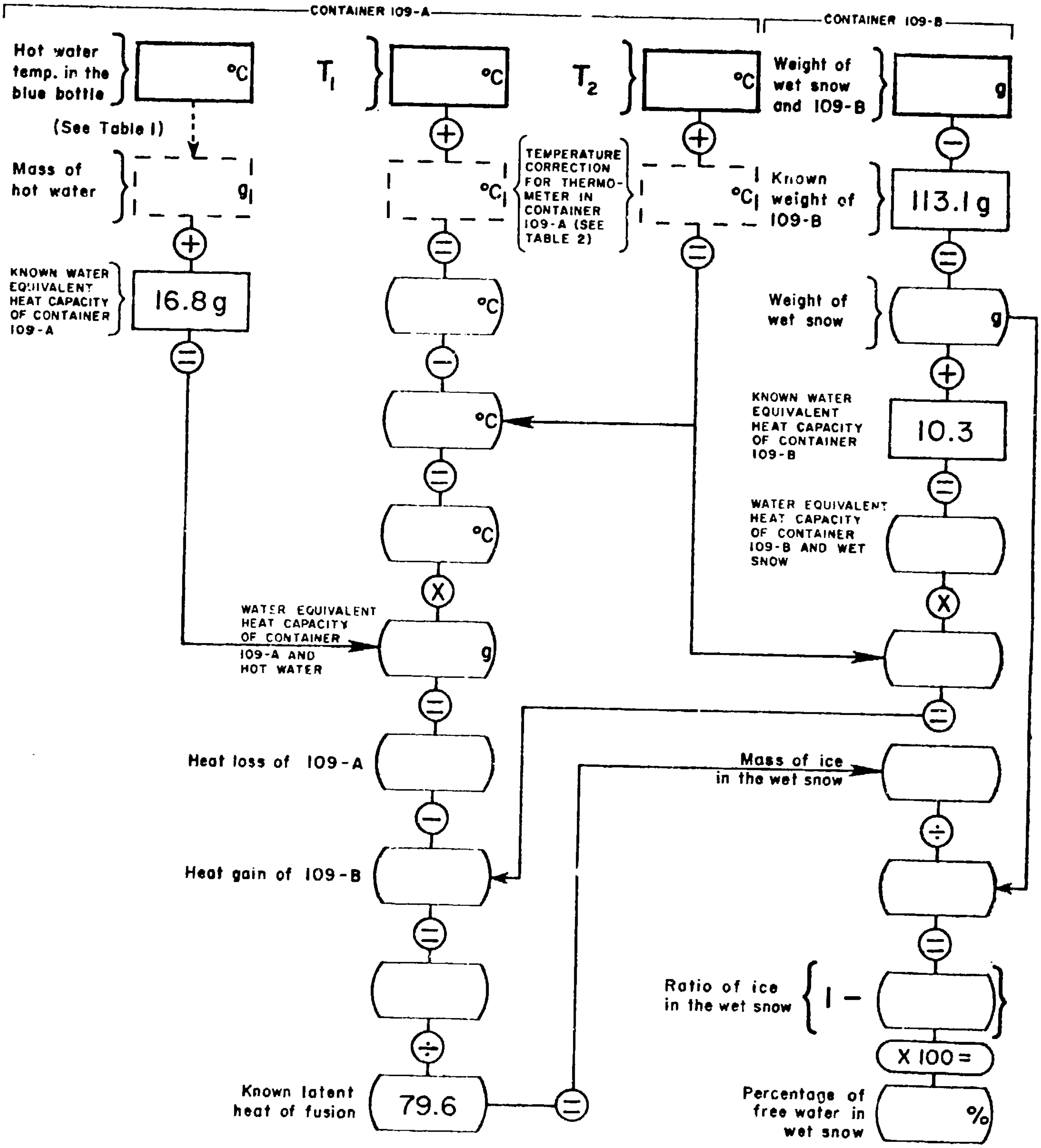
(11) Shake very well.

(12) Read the final temperature of container 109-A.

(13) After reading the final temperature with the thermometer in the container, 109-A, the water in the container should be thrown away and the small amount of water left in the container should be sucked out by a dropper. This readies the calorimeter for its next use.

FREE WATER CONTENT CALCULATION CHART

Measurement Number	Remarks		
Survey Point	Weather	Wind Direction	Container Number
Date	Air Temperature	Wind Velocity	Surveyor's Name



CALORIMETER CONVERSION
TABLE I

Temperature (°C)	Mass (g)
45	251.1
46	251.0
47	250.9
48	250.8
49	250.6
50	250.5
51	250.4
52	250.3
53	250.2
54	250.1
55	250.0
56	249.9
57	249.8
58	249.7
59	249.6
60	249.4
61	249.3
62	249.2
63	249.1
64	249.0
65	248.9

Temperature correction for Calorimeter thermometer in the container 109-A

T_1 and T_2	correction value
10-31 °C	- 0.4 °C
31-41	0.3
41-48	0.2
48-50	0.3

T_1: Initial temperature of hot water in the container 109-A

T_2: Final temperature of water in the container 109-A

Water equivalent heat capacity of 109-A is 13.8 gram.

Weight of container 109-B (dead weight) is 113.1 gram.

Water equivalent heat capacity of container 109-B is 10.3 gram.

APPENDIX G

STRATIGRAPHIC PROFILES at SITES 10B, 16J, 9J, & 8B

Site 10B (elev. 3500 feet) 17h30m 11 Aug., 1972

Juneau Icefield (Measured by G. Wakahama, Y. Endo & H. Nishio)

depth (cms)	snow density (g/cm³)	grain size	Hardness depth (cm)	Hardness kg/cm²	Free water content depth (cm)	Free water content %
surface	0.57	C.d	0	3.1	0-10	18.4
15-22	0.57	C.d				
40-4[illegible]	0.55	C				
81-93	0.58	C	81	[illegible].0	75-80	7.3
117-121	0.52	C	113	2.7		
141-150	0.59	C	141	5.7		
205-210	0.58	C-d			165-170	7.1
230-235	0.[illegible]9	C-d	235	2.9	225-230	4.9
250-255	0.[illegible]	d				
290-295	0.75	C-d	290	2.3	285-290	5.7
295-301	0.75	C-d				
350-355	0.62	C-d	345	4.4	305-310	11.9
390-395	0.66	C-d	390	3.8	350-360	9.1
425-430	0.[illegible]4	C-d	425	14.0	430-440	3.7
447-455	0.72	C-d	447	5.8	485-495	4.6
477-482	0.77	C-d				

Air Temperature		Snow/firn Temperature	
18^h30^m	+2.9°C	surface	0°C
19 15	+3.5	296 cm	0
19 40	+2.8	325	0
20 00	+2.8	400	0
		495	0

Site C-16 junction (elevation 4100 feet) 21h30m-23h

Juneau Icefield 11 Aug., 1972

depth (cms)	snow density (g/cm3)	grain size	Hardness depth (cms)	Hardness kg/cm²	Free water content depth (cms)	Free water content %
surface	0.52	C.d	0	0.41	10-20	3.1
50-55	0.59	C.d	50	3.1	60-70	5.8
90-95	0.58	C.d			130-140	4.7
135-140	0.[illegible]	C.d	135	4.8		
153-158	0.70	C.d	153	0.8	200-210	6.4
190-195	0.70	C.d	19[illegible]	1.1	260-270	5.7
270-275	0.[illegible]9	C.d	28[illegible]	3.1	312-320	6.2

Air Temperature	
21^h50^m	+0.8°C
22 00	+1.0

Site half mile south of C-9 junction 12h - 14h

12 Aug.,1972 Juneau Icefield (elevation 4900 feet)

depth (cms)	density g/cm³	grain size	Hardness depth (cms)	Hardness kg/cm²	Free water content depth (cms)	Free water content %
surface	0.56	d	0	1.1	0-10	15.0
70-75	0.64	c	70	5.0	15-20	7.4
140-145	0.64	c	140	10.0	70-80	7.4
190-195	0.66	c	190	3.0	150-160	6.5
230-235	0.73	c	220	2.1	215-230	13.6
250-255	0.66	c	250	4.1	260-265	9.4
310-315	0.82	c	295	0.8	320-325	9.2
340-345	0.74	c-d	340	2.4	350-360	10.0
380-385	0.63	c-d	380	4.2		
425-430	0.64	c-d	425	16.0	465-475	3.8
455-460	0.66	c-d				

Air Temperature

12^h20^m	+2.8°C
13 00	+2.8

Site C8/C18 junction 16h10m-18h 12 Aug., 1972

Juneau Icefield (elevation 5850 feet)

depth (cms)	density g/cm³	grain size	Hardness depth (cms)	Hardness kg/cm²	Free water content depth (cms)	Free water content %
surface	0.53	c-d	0	0.28	10	11.3
70-75	0.71	c-d	70	1.5	60-65	11.6
130-135	0.64	c-d	130	0.6	140-150	9.7
195-200	0.59	c-d	200	4.6	180-190	5.4
245-250	0.62	c-d	245	8.4	250-260	4.8
290-295	0.68	c-d	290	3.5	295-305	10.5
350-355	0.59	c-d	350	7.4	320-330	8.7
380-385	0.56	d	380	3.2	385-395	5.3
410-415	0.62	d	410	0.6	445-455	6.9
430-435	0.81	c	440	1.6		
490-495	0.78	c	490	.23	480-490	12.0

Air Temperature

18^h30^m	+4.0°C
17 40	+4.3